Ontario Geological Survey
Miscellaneous Paper 123

Geology and Petrogenesis of the Archean Abitibi Belt in the Kirkland Lake Area, Ontario

L.S. Jensen and F.F. Langford

1985

Ontario
Ministry of
Natural
Resources

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Printed in Ontario, Canada

ISSN 0704-2752
ISBN 0-7729-0145-7

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Jensen, L.S., and Langford, F.F.
1985: Geology and Petrogenesis of the Archean Abitibi Belt in the Kirkland Lake Area, Ontario; Ontario Geological Survey, Miscellaneous Paper 123, 130p. Accompanied by Maps P.2434 and P.2435, scale 1:63 360 or 1 inch to 1 mile and sheet of microfiche.

Scientific Editor: Guy Kendrick
1000-85-U.of T.

Foreword

The Abitibi Belt is the most extensive Archean "greenstone" belt in the world and it is also one of the best preserved "greenstone" belts in terms of its relatively low metamorphic grade and structural complexity. The Abitibi Belt presents a unique opportunity to study and to unravel geological problems involved with the development of the Earth's crust during the Early Precambrian.

The author has mapped in the Kirkland Lake area of the Abitibi Belt for the last 15 years. Drawing from his experience in that very important mining area, L.S. Jensen has completed a stratigraphic and petrogenetic synthesis which forms the basis of both his Ph.D. thesis and this volume.

As well as providing a complete and up-to-date summary of the stratigraphy and tectonic history of the eastern part of the Abitibi Belt, this report discusses many new and thought-provoking concepts regarding crustal and "greenstone" belt evolution that have far-reaching implications for Archean metallogenesis.

Hence, it is expected that this work will prove to be a major contribution to our understanding of not only the Abitibi Belt, but also of the geological processes which operated worldwide prior to 2600 Ma ago.

V.G. Milne
Director
Ontario Geological Survey

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GEOLOGICAL MAPS (back pocket)

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Map P.2434 (uncoloured) - Precambrian Geology of the Magusi River Area, Districts of Cochrane and Timiskaming, scale 1:63 360 or 1 inch to 1 mile.

APPENDICES ON MICROFICHE (back pocket)

Appendix 1 - List of chemically analyzed rock (specimens) from the Lightning River map area.

Appendix 2 - List of chemically analyzed rock (specimens) from the Magusi River map area.

CONVERSION FACTORS FOR MEASUREMENTS IN ONTARIO GEOLOGICAL SURVEY PUBLICATIONS

If the reader wishes to convert imperial units to SI (metric) units or SI units to imperial units the following multipliers should be used:

CONVERSION FROM SI TO IMPERIAL			CONVERSION FROM IMPERIAL TO SI		
SI Unit	*Multiplied by*	*Gives*	*Imperial Unit*	*Multiplied by*	*Gives*
			LENGTH		
1 mm	0.039 37	inches	1 inch	**25.4**	mm
1 cm	0.393 70	inches	1 inch	**2.54**	cm
1 m	3.280 84	feet	1 foot	**0.304 8**	m
1 m	0.049 709 7	chains	1 chain	20.116 8	m
1 km	0.621 371	miles (statute)	1 mile (statute)	**1.609 344**	km
			AREA		
1 cm²	0.155 0	square inches	1 square inch	**6.451 6**	cm²
1 m²	10.763 9	square feet	1 square foot	**0.092 903 04**	m²
1 km²	0.386 10	square miles	1 square mile	2.589 988	km²
1 ha	2.471 054	acres	1 acre	0.404 685 6	ha
			VOLUME		
1 cm³	0.061 02	cubic inches	1 cubic inch	**16.387 064**	cm³
1 m³	35.314 7	cubic feet	1 cubic foot	0.028 316 85	m³
1 m³	1.308 0	cubic yards	1 cubic yard	0.764 555	m³
			CAPACITY		
1 L	1.759 755	pints	1 pint	0.568 261	L
1 L	0.879 877	quarts	1 quart	1.136 522	L
1 L	0.219 969	gallons	1 gallon	**4.546 090**	L
			MASS		
1 g	0.035 273 96	ounces (avdp)	1 ounce (avdp)	28.349 523	g
1 g	0.032 150 75	ounces (troy)	1 ounce (troy)	**31.103 476 8**	g
1 kg	2.204 62	pounds (avdp)	1 pound (avdp)	**0.453 592 37**	kg
1 kg	0.001 102 3	tons (short)	1 ton (short)	**907.184 74**	kg
1 t	1.102 311	tons (short)	1 ton (short)	**0.907 184 74**	t
1 kg	0.000 984 21	tons (long)	1 ton (long)	**1016.046 908 8**	kg
1 t	0.984 206 5	tons (long)	1 ton (long)	**1.016 046 908 8**	t
			CONCENTRATION		
1 g/t	0.029 166 6	ounce (troy)/ ton (short)	1 ounce (troy)/ ton (short)	34.285 714 2	g/t
1 g/t	0.583 333 33	pennyweights/ ton (short)	1 pennyweight/ ton (short)	1.714 285 7	g/t

OTHER USEFUL CONVERSION FACTORS

1 ounce (troy)/ton (short)	20.0	pennyweights/ton (short)
1 pennyweight/ton (short)	0.05	ounce (troy)/ton (short)

NOTE–Conversion factors which are in bold type are exact. The conversion factors have been taken from or have been derived from factors given in the Metric Practice Guide for the Canadian Mining and Metallurgical Industries published by The Mining Association of Canada in co-operation with the Coal Association of Canada.

Geology and Petrogenesis of the Archean Abitibi Belt in the Kirkland Lake Area, Ontario

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Manuscript approved for publication by John Wood, Chief Geologist, Ontario Geological Survey, May 2, 1983. Published with the permission of V.G. Milne, Director, Ontario Geological Survey.

Abstract

The volcanic sequence in the Kirkland Lake part of the Abitibi "Greenstone" Belt consists of up to 3, as yet unnamed, supergroups, each comprising a single volcanic cycle. The uppermost supergroup is preserved in a large eastward-plunging synclinorium bounded to the north by the Destor - Porcupine Fault Zone and to the south by the Kirkland Lake - Larder Lake Fault Zone. The synclinorium contains the komatiitic Stoughton - Roquemaure and Larder Lake Groups overlain in turn by the tholeiitic Kinojevis Group and calc-alkalic Blake River Group, each group is about 10 000 m thick. In several graben along the fault zones, the Blake River volcanic rocks were succeeded by alkalic volcanic rocks and fluvial sediments (Timiskaming) and finally, by potassium-rich granitic intrusions.

Found on the outer parts of the synclinorium are separate lower supergroups of about the same age. To the south are the komatiitic Wabewawa Group, tholeiitic Catherine Group, and calc-alkalic Skead Group of one supergroup overlying a still older supergroup represented by the calc-alkalic volcanic rocks of the Pacaud Tuffs intruded by the Round Lake Batholith. To the north is the calc-alkalic Hunter Mine Group cut by the Lake Abitibi Batholith and to the west, at the nose of the synclinorium, are the komatiitic, tholeiitic, and calc-alkalic volcanic rocks of the Deloro Group truncated by the Kenogamissi Batholith. Some trondhjemitic rock was also developed in the calc-alkalic core (Blake River Group) of the synclinorium. Each volcanic cycle is thought to show the primary phase of sialic crustal formation. Where the calc-alkalic and sedimentary rocks are largely replaced by granitic rocks, the sima-to-sial tranformation was nearly completed.

'Megacauldrons', more than 100 km across, represent both the locations and the processes by which primary crust is converted into the continental crust of Archaen shields. In each megacauldron, subsidence of the crustal rocks was initiated by a thick mafic crust of komatiitic and magnesium-rich tholeiitic lavas forming above a diapir originating from deep within an undepleted carbonaceous chondritic mantle. Development of shallow magma chambers in the sinking lavas allowed the fractionation and the extrusion of iron-rich tholeiitic magmas. At depth in the megacauldron, the mafic lavas were converted to garnet-bearing amphibolites and eclogites and were partly melted to form calc-alkalic magmas which migrated to the surface to form calc-alkalic volcanic piles. The resulting heavy residue descended further into the mantle. Likewise, subsequent subsidence of the calc-alkalic volcanic rocks resulted in granitic magmas forming batholiths at the surface and the obliteration of the pre-existing volcanic rocks.

In the youngest megacauldron of the Kirkland Lake area, although formation of granitic rocks did begin, they are not extensive. Hence, most of the volcanic rocks remain so that the different stages of the sima-to-sial transformation in the youngest megacauldron can still be observed. This model can be used as a standard when studying other greenstone belts with less preserved volcanic successions in Canada, Africa, and Australia. In these other greenstone belts, the ultramafic and mafic volcanic rocks were preserved between megacauldrons because some of the mafic volcanic rocks of the younger adjacent and proximal megacauldrons were deposited on sedimentary and felsic calc-alkalic rocks near the margin of a more advanced primary megacauldron. Buoyed by these lighter rocks, the mafic and ultramafic rocks escaped burial and sinking to become surrounded by granitic rocks.

The low metamorphic grade throughout the region indicates that rocks were buried to a maximum of 3 km. The lack of penetrative deformation suggests that the structures formed mostly by subsidence during volcanism.

Geology and Petrogenesis of the Archean Abitibi Belt in the Kirkland Lake Area, Ontario, by L.S. Jensen and F.F. Langford. Ontario Geological Survey, Miscellaneous Paper 123, 130p. Published 1985. ISBN 0-7729-0145-7.

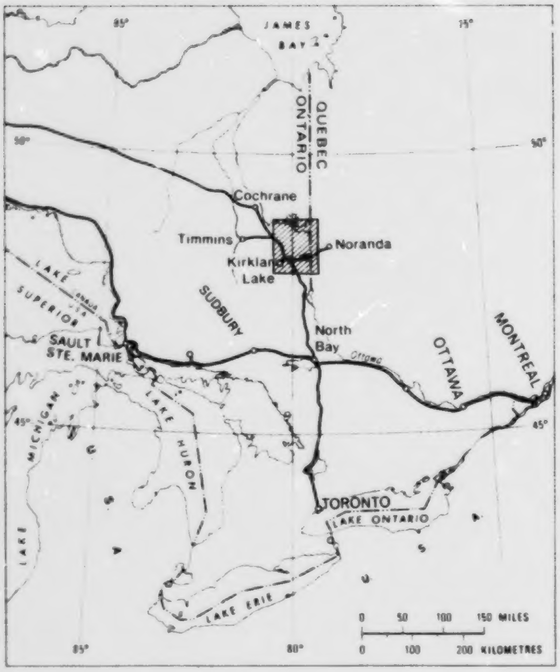


Figure 1. Key map showing location of the Kirkland Lake area.

Introduction

PURPOSE

The purpose of this study was to build a model for the development of Archean greenstone belts by investigating the petrography, chemistry, and field relationships of the Archean rocks of the Abitibi Greenstone Belt in the Kirkland Lake area. The establishment of such a model is especially desirable because Archean greenstone terrains present only a 2-dimensional picture of the earth's Archean crust in its final stage of development. The search for economic deposits might be facilitated if such a model predicted subsurface trends and facies changes.

The Kirkland Lake part of the Abitibi Belt has a volcanic sequence 50 km thick, consisting of 2 successive ultramafic to felsic volcanic cycles. In the upper volcanic cycle, the basal sequence comprises komatiitic lavas that are 10 km thick. The basal sequence overlies calc-alkalic rocks of the lower volcanic cycle. Such a sequence appears to be equivalent to many volcanic successions found in greenstone belts where komatiites have been recognized.

A 10 km thick succession of tholeiitic lavas showing a strong upward iron enrichment lies above the komatiitic succession. This tholeiitic succession contains only tholeiitic rocks and does not appear to be duplicated in other well described greenstone belts. It has some features which are similar to ocean floor tholeiitic successions.

A 10 km thick succession of calc-alkalic basalts, andesites, dacites, and rhyolites lies above the iron-enriched tholeiitic succession. Above the calc-alkalic succession, is a succession of alkalic volcanic rocks that are about 3 km in thickness. Neither the calc-alkalic nor the alkalic successions appear to have equivalent counterparts in other greenstone volcanic sequences. These successions, however, have some features which are similar to calc- alkalic and alkalic successions found in island-arc environments.

The occurrence of thick komatiitic, tholeiitic, calc-alkalic, and alkalic volcanic successions in one volcanic sequence suggests that the Kirkland Lake area contains a more complete volcanic sequence than other Archean greenstone belts. Therefore, this area should be ideal for formulating a model to account for the development of Archean greenstone belts.

STATEMENT OF THE PROBLEM

Many divergent opinions are held about how Archean crustal rock developed between approximately 3.8 and 2.5 Ma ago, mostly because of ambiguous field evidence (Moorbath 1975a; Windley 1976, 1979). Fundamentally, Archean crustal rock consists of linear to arcuate volcano-sedimentary belts engulfed by granitic rocks (Goodwin 1968; Anhaeusser *et al.* 1969; Glikson 1972; Windley and Bridgwater 1971; Moorbath 1976). To account for this feature, models for the development of Archean crustal terrains have been developed, based on either of 2 opposing hypotheses: (1) sima-to-sial transformation, and (2) remobilized sialic crust.

The hypothesis of processes involving transformation of sima to sial that dominated during the Archean has been favoured by: Arth and Hanson (1972, 1975), Engel (1968), Viljoen and Viljoen (1969a), Glikson (1972, 1976a, 1976b, 1976d), Glikson and Sheraton (1972), Glikson and Lambert (1976), Moorbath (1975a, 1975b, 1976, 1977), Anhaeusser (1971a, 1973, 1975a), O'Nions and Pankhurst (1974), Condie and Hunter (1976), Condie (1975), and Hanson (1978). The other hypothesis suggests that much of the exposed granitic rock is the result of the anatexis of an earlier formed sialic crust in and on which volcanic and sedimentary rocks were formed and later deformed during the anatexis (MacGregor 1951; Ramberg 1964; Green 1972a; Fyfe 1973a, 1973b, 1978a, 1978b; Sutton 1971, 1973; Hargraves 1976; Baragar and McGlynn 1976; Shaw 1976; Collerson and Fryer 1978; Bridgwater and Fyfe 1974; Taylor and Hallberg 1977).

Stemming from the 2 hypotheses above, several models for the formation of Archean greenstone belts have been proposed (Windley 1977a). Some models favour a plate tectonics similar to that of modern plate tectonics (Engel 1968; White *et al.* 1971; Goodwin and Ridler 1970; Condie and Baragar 1974; Langford and Morin 1976; Condie and Harrison 1976), or an 'accelerated' form of plate tectonics involving rapid mantle convection (Fyfe 1976; Hart *et al.* 1970) which would allow tectonic crustal thickening by thrust faulting (Tarney and Windley 1977; Coward *et al.* 1976; Bridgwater *et al.* 1973). Others have rejected plate tectonics (Hargraves 1976; Glikson and Lambert 1976; Shaw 1976; Lambert 1976; Taylor 1979) in favour of vertical tectonics involving gravitational forces (MacGregor 1951; Windley and Bridgwater 1971; and Baragar and McGlynn 1976). The latter models would allow greenstone belts to develop from volcanic plumes (Hunter 1974; Condie and Hunter 1976; Condie 1975; Collerson and Fryer 1978; Fyfe 1978b) which may have formed in rifts occurring in the earlier sialic crust (Anhaeusser 1973) or by meteoritic bombardment (Donn *et al.* 1965; D.H. Green 1972a, 1972b, 1975; Condie and Baragar 1974). In addition, crustal thickening could have taken place by magmatic underplating (Holland and Lambert 1976; Fyfe 1978b).

None of these models have won general acceptance. The main reason for this is that there is either insufficient field evidence to support these models or that they fail to explain adequately the generation of the rock types or certain features unique to other Archean greenstone belts. This is probably a result of the fact that the models are based on greenstone belts whose volcanic successions are largely truncated by granitic intrusions.

Critical to any model is the nature of the pre-existing crustal material and the conditions which led to, and governed the formation of Archean crustal terrains. Some models have been based on the assumption that the geothermal gradients were 2 to 3 times higher (40 to 60°C/km) during the Archean period and that only a thin crust was possible (Hart *et al.* 1970; Windley and Bridgwater 1971; Fyfe 1973a, 1973b, 1978a, 1978b; Shaw 1976; Sinha 1976; O'Hara 1977; Lambert 1976; Collerson and Fryer 1978; Baer 1977). Others have presented evidence that the Archean crust was as thick or thicker than at present (Arth and Hansen 1972, 1975; Green 1972b; Condie 1973, 1976a; Moorbath 1975a; Cawthorn 1975; Wells 1976; Glikson and Lambert 1976; Windley 1977b; O'Hara 1977; Tarney and Windley 1977; Windley and Davies 1978; Sun and Nesbitt 1978a; Taylor 1979). This implies the presence of a gradient of 30°C/km and one even as low as 15°C/km (Wells 1976) during the Archean.

The thin Archean crustal models with high geothermal gradients require the Earth to have undergone an earlier molten stage during which a thin worldwide igneous (preferably granitic) crust was formed and there would be, by necessity, a high heat flow to dissipate the heat resulting from cooling and the release of radioactive heat (Fyfe 1973a; Moorbath 1975a; Shaw 1976; Hargraves 1976; Sinha 1976; Collerson and Fryer 1978). The molten stage of the Earth would have resulted mainly from accretional energy and the release of energy during the segregation of the Earth's material into a core, mantle, and crust (Shaw 1976). This could have occurred if the Earth accreted homogeneously from a solar nebula, as proposed by Hanks and Anderson (1969).

On the other hand, the Earth may have escaped a molten stage if (a) the Earth accreted heterogeneously from a differentiated solar nebula as proposed by Grossman (1972); and (b) it was a 'cold' accretion (Brown 1952; Urey 1957; Dietz 1965; Fisher 1976). If this had occurred, the core material could have accreted first, and segregated into a molten core and lower mantle, to be covered later by a veneer of cold volatile-rich nebular condensate (Hutchinson 1974; Fisher 1976; Walker 1976; Murthy 1976). Fisher (1976) presented evidence that degassing is still occurring, and therefore supports the idea that the Earth has never undergone a molten stage. If this had been the case, the protocrust of the Earth would have been thicker than the Earth's present crust. This would have allowed for a heterogeneous upper mantle with much of it undepleted, to be present during the Archean (Condie 1976a; Sun and Nesbitt 1977), since the protocrust and the upper mantle would have been one and the same (Dietz 1965). More recently, Ringwood (1979) has suggested a homogeneous accretion model whereby a molten core could develop without melting the upper mantle.

Whether the Earth during the pre-Archean had a worldwide thick cool protocrust composed of nebular condensate similar to carbonaceous chondrite, or a thin, hot floating, granitic crust 10 km in thickness, either condition would have greatly affected the subsequent development of Archean greenstone-granite terrains. Therefore, if a model for the development of Archean crustal terrains is to be successful, it must resolve the above controversy and at the same time account for the distribution of rock types as they occur in Archean terrains, particularly in greenstone belts.

In this study, it is proposed that the Archean crust developed by the transformation of simatic material to sialic material within megacauldrons located on a carbonaceous chondrite protocrust. Greenstone belts formed when primary megacauldrons were succeeded by adjacent and subjacent megacauldrons, and represent volcanic and sedimentary rocks which escaped being transformed into sialic crust during the development of the secondary megacauldrons. Subsidence of the volcanic and sedimentary rocks marginal to the granitic cores of the megacauldrons gave the greenstone belts their arcuate synclinal form.

AREA AND LOCATION

This study concentrates on the Archean rocks of the Kirkland Lake area in northeastern Ontario (*see* Figure 1). The general study area occurs within Longitudes 79°30'W and 80°39'W, and Latitudes 48°00'N and 48°45'N, an area of nearly 6200 km^2. Located within the boundaries of the area are the gold mining communities of Kirkland Lake, Larder Lake, Virginiatown, Swastika, and Holtyre, and the farming communities of Matheson and Ramore (Figure 2).

Within the above area, 2 smaller areas were selected for detailed investigation and are the Lightning River Area and the Magusi River Area, both located northeast of Kirkland Lake (Figure 2). These are adjoining areas bounded by Longitudes 79°30'W and 80°00'W, and Latitudes 48°15'N and 48°45'N and together, cover an area of about 2000 km^2.

PREVIOUS WORK

Geological reconnaissance parties were sent into the Kirkland Lake area by the provincial and the federal governments from 1900 to 1940 to delineate areas of volcanic, sedimentary, and granitic rocks and describe "newly" discovered zones of mineralization. By 1920, a simple stratigraphic code had been developed for the area. The code was 'Keewatin' for predominantly volcanic sequences; 'Timiskaming' for folded, predominantly sedimentary sequences containing conglomerate; and 'Algoman' for all the granitic rocks.

In the late 1930s, a program for the detailed mapping of individual townships was initiated by the Ontario Government. This program continued until 1975. In general, no detailed stratigraphic correlations were made except in isolated cases. Attempts were made, however, to extend the early stratigraphic divisions westward from the Rouyn - Harricanaw Region of Quebec, where Gunning and Ambrose (1939) had divided the Early Precambrian rocks (Archean) into 4 conformable groups from the oldest to the youngest as follows:

1. Cadillac Group, chiefly sedimentary rocks
2. Blake River Group, chiefly volcanic rocks
3. Kewagama Group, chiefly sedimentary rocks
4. Malartic Group, chiefly volcanic rocks

In Quebec, Ambrose (1941) extended the above terminology to include the volcanic and sedimentary rocks of the Noranda region. Ambrose placed the volcanic rocks of Arguebelle and Manneville Townships, in Quebec, as well as the volcanic rocks northwest of Noranda into the the Blake River Group. Norman (1944) introduced the term Kinojevis Group for the volcanic rocks in Arguebelle and Manneville Townships because the volcanic rocks in places appeared to underlie the sedimentary rocks of the Kewagama Group. Ambrose (1944) adopted the term Kinojevis Group, and was able to map this group westward into Ontario. There, Ambrose placed the

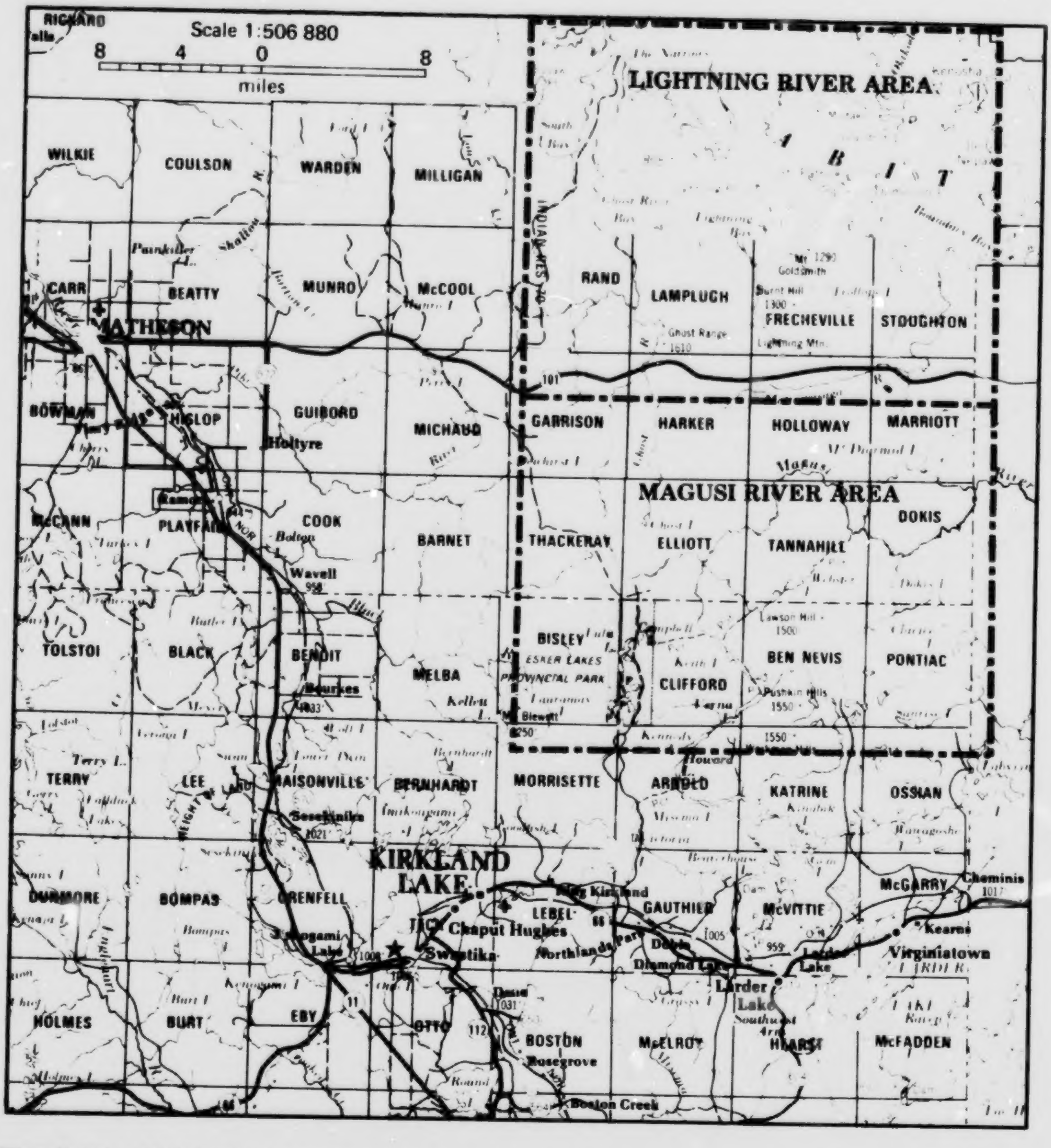


Figure 2. Map of Kirkland Lake area showing the area of detailed mapping.

Blake River Group south of the Destor - Porcupine Fault and the Kinojevis Group north of the fault. Since then, M.E. Wilson (1962), Baragar (1968), and Stockwell *et al.* (1970), have retained the name 'Blake River' Group for the volcanic rocks of Arguebelle and Manneville Townships, while Tremblay (1950), Jones (1964) and Dimroth (1975a, 1975b) have favoured the name 'Kinojevis' Group.

Roscoe (1965) noted that the Blake River Group in Ontario is underlain by a thick basaltic succession that has no counterpart within the Blake River Group in Quebec. Baragar (1968), however, placed this thick basaltic succession in the Blake River Group, noting that it was similar in chemical composition to the volcanic rocks in Arguebelle Township which he also included in the Blake River Group. Dimroth *et al.* (1974) noted that a variolitic marker horizon occurs at the base of the Blake River Group in La Pause Township, Quebec. From there, this marker horizon extends to the interprovincial boundary where it was noted by Baragar (1968) to continue westward through Marriott Township. More recent studies (Goodwin 1977a, 1977b, 1979; Naldrett and Goodwin 1977) have placed all the rocks south of the Destor - Porcupine Fault within the Blake River Group, whereas Jensen (1976a, 1978a, and 1978b) has used the term Kinojevis Group to describe the basalts of Roscoe (1965) that lie below the variolitic marker horizon of

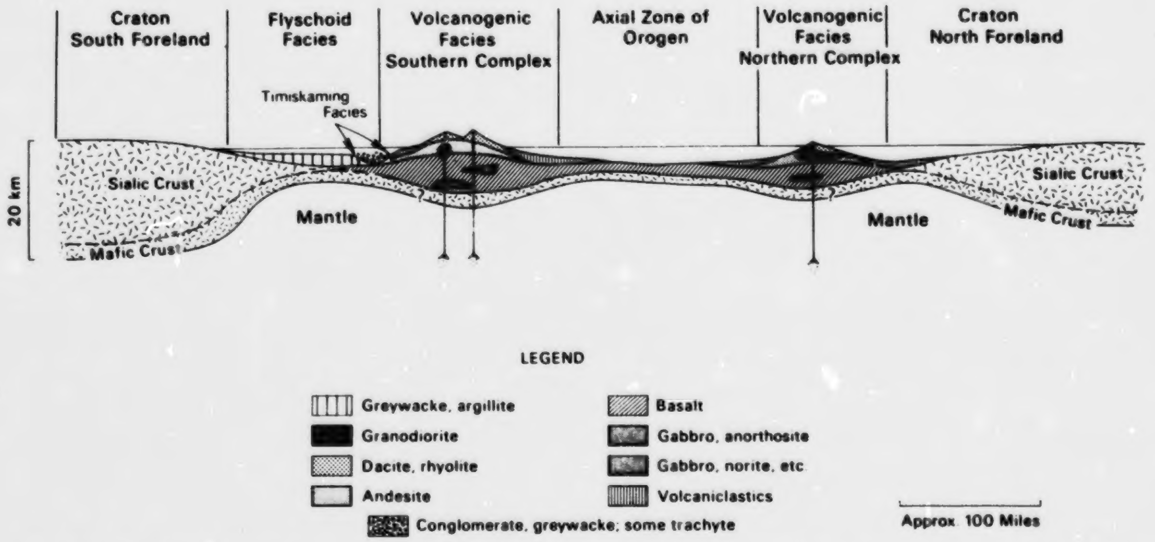


Figure 3. Hypothetical tectonic reconstruction of Abitibi orogen. Vertical cross-section. Length of the orogen is approximately 500 miles. Distance between southern and northern volcanogenic facies may have been in the order of 200 miles. Vertical scale is approximate only. The line of the section is intended to cross Noranda in the south and Matagami in the north. The smaller size and relief of the northern sialic foreland is indicated by the absence of flyschoid facies along its southern margin. The nature and thickness of mafic crust underlying the orogen is conjectural. In the absence of direct evidence of sialic contribution to the volcanic and sedimentary assemblages of the orogen (except for the flyschoid facies, i.e. Pontiac, at the southern margin), no subjacent sialic crust is shown as being present, at least during the early development stages of the orogen (after Goodwin and Ridler 1970).

Dimroth (1975a). It has been generally assumed that the Cadillac Group of Quebec is equivalent to the Timiskaming Group in Ontario, and that the Blake River Group extends westward into Ontario to the north of Kirkland Lake (M.E. Wilson 1962).

During the 1960s, several studies have been carried out on the Archean rocks in the Kirkland Lake area. Bass (1961) proposed that the unconformity between the Keewatin and Timiskaming Groups was local, and that the Archean rocks in this area represented a single orogenic cycle. This interpretation was based on the observation that 'granite' pebbles were a negligible component of Timiskaming-type conglomerates, and proposed further that the felsic pebbles might be derived from subvolcanic intrusions associated with nearby island-arc volcanism. This island-arc concept was used by Cooke (1966), after he had completed a study of the alkalic volcanic rocks in the Kirkland Lake area. He concluded that the alkalic volcanic rocks interlayered with the Timiskaming-type sedimentary rocks represented a late stage in the evolution of island-arc calc-alkalic volcanism, and that the Timiskaming rocks were laid down in a shallow sea on the continental side of an island-arc.

Kalliokoski (1968) questioned correlation of 'conglomerates' across the Larder Lake Break. His doubt was based on the strong lithological differences across the 'break' noted by Thomson (1949) and Hewitt (1963). Kalliokoski (1968) felt that the south side of the 'break' had moved up relative to the north side, and therefore, a simple correlation across the 'break' was incorrect. This view was later supported by Jensen (1975c).

Ridler (1969, 1970) proposed that the 'Larder Lake Break' did not exist. Instead, he claimed that the 'break' represented foliation of the incompetent carbonate rocks in the Timiskaming-type sedimentary rocks. He also suggested that because the volcanic rocks enclosing the Boston iron-formation were relatively high in alkalies, they should be regarded as "Timiskaming"; hence the carbonate rocks were a facies of the iron-formation. Jensen (1974) and Tihor and Crocket (1977) pointed out that many of the carbonate rocks were carbonatized peridotitic komatiites. Jensen (1976b, 1977) also pointed out that sedimentary rocks, mapped as "Timiskaming", south of the 'Larder Lake Break' were much older than those north of the 'break' and that many of Ridler's (1970) alkali-rich volcanic rocks were actually komatiites affected by syenite intrusions.

Based on Ridler (1969, 1970) and additional chemical studies by Goodwin (1965) in the Abitibi Belt, Goodwin and Ridler (1970) postulated that the Abitibi Belt resulted from intercratonic ocean- floor spreading which occurred in a manner similar to modern plate tectonics. This spreading enabled mafic to felsic volcanic sequences to develop near the sialic forelands (Figure 3). Following this, the volcanic rocks would be affected by regional tectonic deformation and were uplifted and eroded. More recently, Ridler (1975) has also proposed that the large batholiths, for example the Round Lake Batholith, represent remobilized pre-existing cratonic masses.

Hyde (1978) has shown that the Timiskaming conglomerates near Kirkland Lake were deposited in the channels of rivers, and the conglomerates near

Larder Lake were deposited in a marine environment. He considered the 2 types of sedimentary rocks to be related to each other because of their proximity, although he was unable to locate sediments which could be ascribed to a shallow marine or shelf environment. He suggested that the lack of shelf sediments indicated a volcanic island environment.

The alternative view, that the 2 types of sedimentary rock are not correlative (Kalliokoski 1968; Jensen 1976b, 1977), would explain the lack of shelf sediments and lend weight to the view that the 'Larder Lake Break' exists (Thomson 1949).

Baragar (1968) conducted a chemical study on the volcanic rocks along the Ontario - Quebec boundary northeast of Kirkland Lake. He proposed that the thick basaltic succession of tholeiites noted by Roscoe (1965) was derived by partial melting of the mantle at a depth of about 60 km. Baragar (1968) suggested further that the overlying high alumina basalts and rhyolites were caused by the melting of sialic crust that had been downwarped by the thick pile of tholeiites.

Jolly (1974, 1975, 1976, 1978, 1980) carried out a broad sampling program that included the Kirkland Lake area in order to study the chemistry and metamorphism of the rocks in the Abitibi Belt. From his findings, Jolly (1975) suggested that the volcanic rocks of the Abitibi Belt could be subdivided into a magnesian suite, a tholeiitic suite, and a primitive calc-alkalic suite. Each of the 3 suites represent different degrees of partial melting of the mantle, followed by near surface crystal-liquid fractionation of the magmas. He also suggested that the magnesian suite occurred at the lowest stratigraphic levels and the primitive calc-alkalic suite at the highest levels.

Numerous studies have been carried out on the ultramafic rocks in the vicinity of Munro Township (MacRae 1969; Pyke *et al.* 1973; Fleet and MacRae 1975; Arndt 1975, 1976; Arndt *et al.* 1977; Whitford and Arndt 1978; Arth *et al.* 1977). Much of this work has been done to document the petrography and chemistry of the komatiitic volcanic rocks. Arndt (1975) suggested that the komatiites were derived from a rising mantle diapir.

SCOPE OF INVESTIGATION

The field mapping was carried out in conjunction with a geological field mapping program for the Ontario Geological Survey. The author was commissioned to make a series of compilation maps 15′ Latitude by 30′ Longitude, each comprising an area of about 100 km^2. Two of these maps, the Lightning River area and the Magusi River area, form the basis of this study. Four other report areas, namely the Ramore, Matheson, Kirkland Lake, and Larder Lake areas, are being completed (Jensen 1974, 1975c, 1976b, 1977, 1978c, 1979).

Field work for the Lightning River and Magusi River areas was completed during the summer field seasons of 1968 to 1973. Mapping was carried out at a scale of 4 inches to 1 mile (1:15 840). Geological information obtained from pace - and - compass traverses was plotted on air photographs and transferred to base maps of the same scale. Traverses were made to all outcrops observed on air photographs, or, were made at intervals of 1/4 mile.

During the field mapping, hand specimens were collected from each outcrop according to the rock types present. In addition, diamond- drill core samples were collected where possible. From the 10 780 hand samples and diamond-drill core samples, thin sections were cut from 1531 samples and analyses for the major elements from 729 samples.

All of the chemical analyses were made by the Geoscience Laboratories of the Ontario Geological Survey, Toronto.

Care was taken in the field to obtain fresh unweathered samples for analysis. Nearly all the samples consisted of 1 to 2 kg of unweathered rock chips (3 mm to 6 cm in maximum dimension). Primary crushing in a Denver crusher reduced the chips to approximately 1.25 cm. Further crushing in a Braun pulverizer with ceramic plates brought the particle size down to less than 10 mesh. Grinding was done in a Fretsch pulverisette 5, planetary ball mill using sintered corundum (11.7% Al_2O_3) balls and vials. The entire sample was reduced to less than 100 mesh. The powder was mixed with lithium tetraborate ($Li_2B_4O_7$) as a flux and lanthanum oxide (La_2O_3) as a heavy absorber; the mixture was in the ratio 1:0, and 6:5. The mixture was heated in a flat bottomed graphite crucible at 1090°C for 15 minutes; the bottom of the resultant tablet was polished with silicon carbide powder and subsequently washed with water and then alcohol in an ultrasonic vibrator.

The tablets were then analyzed on a computer controlled ARL (Applied Research Laboratories) X-ray spectrograph (Model 72000), for SiO_2, Al_2O_3, Fe_2O_3, MgO, CaO, K_2O, Na_2O, P_2O_5, TiO_2. The intensities of the above elements were compared to the intensities of standards prepared in the same manner. Cu, Ni, Zn, and Cr were determined by atomic absorption.

The thin sections were made by the University of Saskatchewan and the Geoscience Laboratories of the Ontario Geological Survey. Microscopic examination of the mineralogy and texture of the rocks was valuable to check the reliability of the chemical analyses and chemical classifications as well as the reliability of the field work. Considerable efforts were made to identify the alteration and metamorphic index minerals.

ACKNOWLEDGMENTS

The author wishes to thank the Ontario Geological Survey, particularly Drs. E.G. Pye (formerly Director) and V.G. Milne (formerly, Chief Geologist), who allowed the author the extra time and effort necessary to investigate many problems found within the map area which are not normally investigated during normal geological mapping. Without the support of the Ontario Geological Survey, such an undertaking would not have been possible by the author alone.

The author also wishes to acknowledge the many fruitful discussions with other members of the staff of the Ontario Geological Survey, particularly, D.R. Pyke and H.L. Lovell. Many fruitful discussions with geologists employed by the mining industry and academia were also beneficial.

Particular thanks are extended to R. Laakso and D. Moddle of the Geoscience Laboratories, Ontario Geological Survey, for their strong laboratory support. Many of the problems would have remained unsolved without their special effort to do the many chemical analyses required for this study.

Special thanks are also extended to F.F. Langford of the University of Saskatchewan who acted as advisor for this study which forms the basis of the senior author's Ph.D. thesis at the University of Saskatchewan. His advice and criticism were valuable assets and required that he be a joint author of this study.

Lastly, thanks are extended to the many field assistants who served on the various field parties directed by the author. Of special mention are D. Hunt, W. Waycheson, T. Hawkins-Clarke, D. Degris, and C. Storey, all of whom were responsible for part of the mapping.

This study is dedicated to my wife Mariel and to my daughters Sylvie and Sonyia for their patience and understanding.

General Geology

REGIONAL SETTING

The Kirkland Lake area is underlain by Archean volcanic, sedimentary, and intrusive rocks which are part of the Abitibi Greenstone Belt. The Abitibi Greenstone Belt is approximately 800 km long and 240 km wide (Figure 4). In places, the belt is cut by swarms of north-trending Matachewan dikes. In the southern part of the Kirkland Lake area, the Archean rocks are unconformably overlain by Proterozoic sedimentary rocks of the Cobalt Group. Several diabase dikes of Keweenawan age cut all the above rocks. Pleistocene deposits consist of till, sand, gravel, and clay which were mostly deposited during the retreat of the Wisconsin glacier (Boissonneau 1966). Recent deposits consist of alluvium and peat.

In the Kirkland Lake area, the volcanic rocks of the Abitibi Greenstone Belt are preserved in a synclinorium located between the Lake Abitibi Batholith and the Round Lake Batholith (Figure 5). The axis of the synclinorium occurs midway between the 2 batholiths and plunges to the east (Jensen 1975a). The northern and southern limbs of the synclinorium are cut by 2 large east-striking fault zones: the Destor - Porcupine Fault Zone, and the Kirkland Lake - Larder Lake Fault Zone, respectively. The 2 fault zones divide the area into 3 large fault blocks with smaller fault blocks occurring in the fault zones.

The Archean rocks of the area have been affected by subgreenschist regional metamorphism (Jolly 1974, 1978).

STRATIGRAPHIC UNITS

The volcanic rocks were formed during cycles of volcanism associated with sedimentation and plutonism. Each cycle consisted of komatiitic volcanism followed by tholeiitic, calc-alkalic, and finally, alkalic volcanism (Jensen 1979). Rocks of the older volcanic cycles are restricted to the outer parts of the synclinorium near the margins of the Lake Abitibi Batholith and the Round Lake Batholith, and are referred to as the Lower Supergroups (Figure 5). Rocks of the younger volcanic cycle form most of the synclinorium and are referred to as the Upper Supergroup. The stratigraphic succession is shown in Figure 5.

The Lower Supergroup on the southern limb of the synclinorium, southeast of Kirkland Lake, contains a cycle of komatiitic, tholeiitic, and calc-alkalic volcanic rocks represented by the Wabewawa, Catherine, and Skead Groups respectively (Jensen 1979). Studies have been made on these rocks by Goodwin (1965), Goodwin and Ridler (1970), and Ridler (1969, 1970, 1975). A trachyte-pebble conglomerate at the top of the Skead Group, suggests that a phase of alkalic volcanism may have occurred at the end of the cycle of volcanism. Below this cycle the calc-alkalic volcanic rocks of the Pacaud Tuffs (Goodwin 1965) suggest that an even older volcanic cycle was developed in the area.

The volcanic cycle represented by the Upper Supergroup consists of komatiitic lavas of the Larder Lake Group succeeded by tholeiitic rocks of the Kinojevis Group and calc-alkalic rocks of the Blake River Group. Alkalic volcanic rocks of the Timiskam-

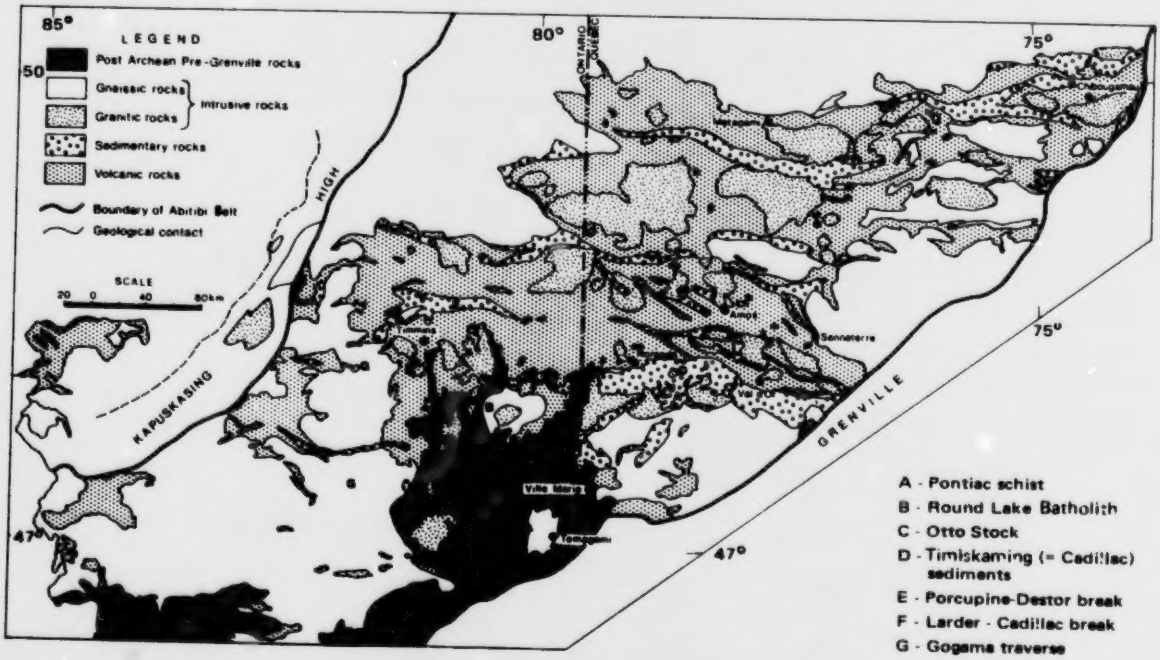


Figure 4. Geological sketch map of the Abitibi Belt simplified from maps of the Quebec Department of Natural Resources and Ontario Geological Survey. The gneissic terrains contain undesignated granitic intrusive rocks equivalent to those shown within the volcanic belts (modified after Jolly 1978).

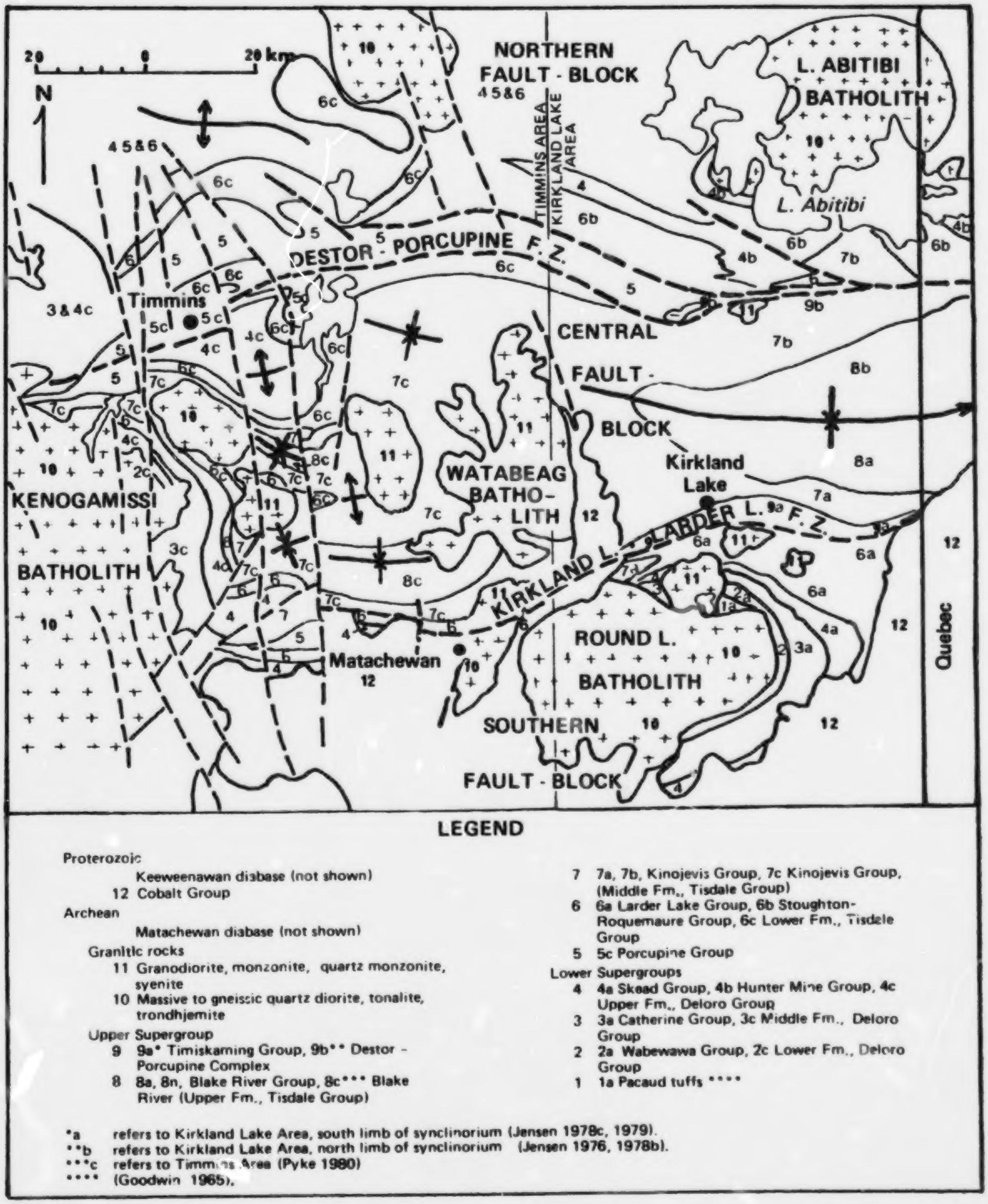
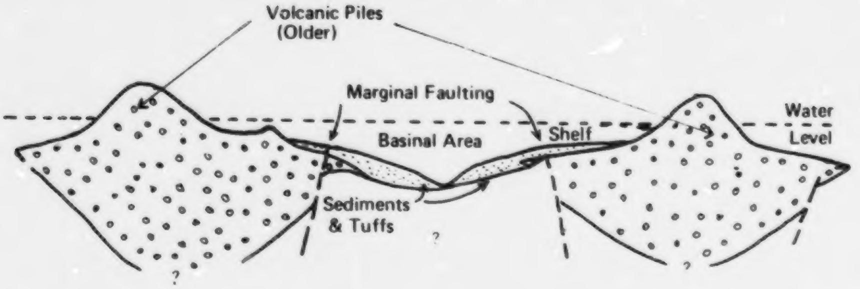
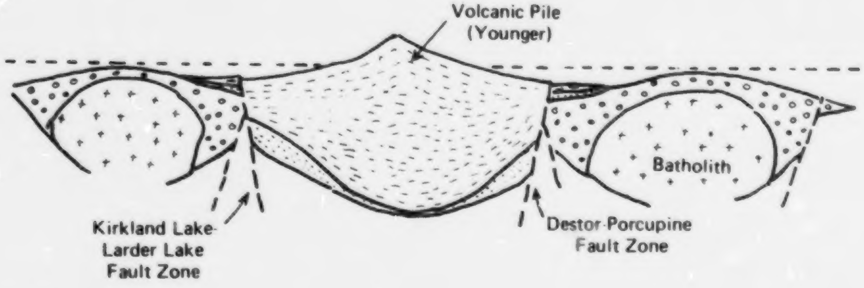

20 0 20 km
N
NORTHERN FAULT - BLOCK
4 5&6
TIMMINS AREA
KIRKLAND LAKE AREA
L. ABITIBI BATHOLITH
L. Abitibi
DESTOR - PORCUPINE F. Z.
Timmins
CENTRAL FAULT - BLOCK
Kirkland Lake
KENOGAMISSI BATHOLITH
WATABEAG BATHOLITH
KIRKLAND L. - LARDER L. F. Z.
ROUND L. BATHOLITH
Matachewan
SOUTHERN FAULT - BLOCK
Quebec
3&4c
LEGEND
Proterozoic
Keeweenawan diabase (not shown)
12 Cobalt Group
Archean
Matachewan diabase (not shown)
Granitic rocks
11 Granodiorite, monzonite, quartz monzonite, syenite
10 Massive to gneissic quartz diorite, tonalite, trondhjemite
Upper Supergroup
9 9a* Timiskaming Group, 9b** Destor - Porcupine Complex
8 8a, 8n, Blake River Group, 8c*** Blake River (Upper Fm., Tisdale Group)
7 7a, 7b, Kinojevis Group, 7c Kinojevis Group, (Middle Fm., Tisdale Group)
6 6a Larder Lake Group, 6b Stoughton-Roquemaure Group, 6c Lower Fm., Tisdale Group
5 5c Porcupine Group
Lower Supergroups
4 4a Skead Group, 4b Hunter Mine Group, 4c Upper Fm., Deloro Group
3 3a Catherine Group, 3c Middle Fm., Deloro Group
2 2a Wabewawa Group, 2c Lower Fm., Deloro Group
1 1a Pacaud tuffs ****
*a refers to Kirkland Lake Area, south limb of synclinorium (Jensen 1978c, 1979).
**b refers to Kirkland Lake Area, north limb of synclinorium (Jensen 1976, 1978b).
***c refers to Timmins Area (Pyke 1980)
**** (Goodwin 1965).

Figure 5. Geological map of the Timmins - Kirkland Lake area.



6a. Basin with marginal volcanic piles



6b. Filling of basin

Figure 6. Configuration of the basin prior to and during the deposition of the younger volcanic cycle (Upper Supergroup).

ing Group unconformably overlie the Blake River and Kinojevis Groups.

On the northern limb of the synclinorium, the oldest rocks are calc-alkalic volcanic rocks of the Hunter Mine Group which are believed to be the top of a volcanic cycle. The lower rocks of this cycle are either destroyed or not exposed. A komatiitic succession, called the Stoughton - Roquemaure Group (Jensen 1976a, 1978b) which forms the basal section of the main volcanic cycle of the Upper Supergroup, unconformably overlies the Hunter Mine Group. The Destor - Porcupine Complex is thought to be analagous to the alkalic volcanic rocks of the Timiskaming Group.

The Larder Lake Group and Stoughton - Roquemaure Group komatiitic successions extend west and join at the nose of the synclinorium in the Timmins area (Jensen and Pyke 1980) where they were referred to as the Lower Formation of the Tisdale Group by Pyke (1980; Figure 5). Rocks of an older volcanic cycle, called the Deloro Group (Pyke 1980), underlie these komatiites. The Hunter Mine, Deloro, Skead, and associated groups probably represented separate but roughly time- equivalent volcanic piles that were formed on the margins of a basin prior to the deposition of the komatiitic rocks belonging to the younger volcanic cycle (Figure 6). Sediments, represented by the Porcupine Group (Pyke 1980), were deposited in the basin at the edges of the Deloro volcanic pile prior to, and during the deposition of the komatiitic rocks of the Tisdale Group (Pyke 1980).

Komatiitic volcanism, at the start of the younger volcanic cycle, occurred throughout the basin and extended onto the margins of the older volcanic piles to form a continuous unit which can now be traced all the way around the synclinorium. The komatiitic volcanic rocks were probably thicker in the central part of the basin than on the edges of the older volcanic piles. Thick successions of tholeiitic and calc-alkalic lavas were also deposited in the basin above the komatiitic succession. For this to have occurred, the original floor of the basin probably sank under the accumulating volcanic rocks to continuously make room for the upper volcanic successions. Jolly (1978) came to a similar conclusion based on the metamorphism of the lavas throughout the area.

The sinking of volcanic rocks during their accumulation would explain the origin of the synclinorium. If the greatest amount of sinking occurred near the centre of the basin, the outer volcanic lavas would tilt and face inward (Figure 6). Likewise, the Destor - Porcupine and Kirkland Lake - Larder Lake Fault Zones probably resulted from differential vertical movement between the older volcanic piles to the north and south and the floor of the basin. Initially, the boundary between the older volcanic piles and the basin would have formed a hinge zone as soon as the basin began to sink. The volcanic strata at the hinge zones would have sheared and would have been displaced downward on the basin side of the fault zones as the basin continued to sink. The tilt of the strata at the same time would have caused tensional forces along the fault zones and allowed fault blocks to drop and create grabens in the fault zones. The grabens would then have been filled with the alkalic volcanic rocks and sediments of the Timiskaming Group.

Trondhjemitic batholiths probably formed in the cores of the older volcanic piles by the partial melting of buried volcanic rock. As the batholiths grew, these batholiths intruded the upper felsic volcanic rocks of the older volcanic cycles, and, in places extended and grew, so that they intruded the komatiitic rocks at the base of the younger volcanic cycle.

GEOCHRONOLOGY OF THE VOLCANIC ROCKS

Nd and Sr isotope work by Zindler *et al*(1978), indicated an age of 2765 ±42 Ma for the komatiites in Munro Township which are considered to be part of the Stoughton - Roquemaure Group. This predates the ages obtained on zircons in the Timmins and Kirkland Lake areas by Nunes and Pyke (1980) and Nunes and Jensen (1980).

South of Timmins, near the Kenogamissi Batholith, the felsic calc- alkalic volcanic rocks of the Upper Formation of the Deloro Group (Pyke 1980) have been dated at 2725 ±2 Ma. These rocks underlie komatiitic rocks correlated with the Stoughton - Roquemaure Group of the Upper Supergroup.

South of Lake Abitibi, the Hunter Mine Group has been dated at 2710 ±2 Ma (Nunes and Jensen 1980). It underlies the Stoughton - Roquemaure Group. This date suggests that the volcanic cycle represented by the Deloro Group is older than the volcanic cycle presented by the Hunter Mine Group.

A preliminary date of 2710 ±2 Ma was obtained on the Skead Group (Nunes Geochronologist, Royal Ontario Museum, personal communication, 1980) at the top of the Lower Supergroup on the southern limb of the synclinorium.

A date of 2703 ±2 Ma was obtained on the felsic volcanic rocks near the top of the Blake River Group of the Upper Supergroup (Nunes and Jensen 1980). This would indicate that the 30 km thick volcanic sequence belonging to the Upper Supergroup on the northern limb of the synclinorium, had an accumulation rate of 4 ±2 cm/year.

REGIONAL METAMORPHISM

Jolly (1974, 1978) concluded that the volcanic rocks in the Kirkland Lake area underwent prehnite - pumpellyite facies metamorphism at pressures of about 1 kb. He also concluded that since prehnite - pumpellyite facies metamorphism occurred on both sides of the Destor - Porcupine Fault, the thick upper part of the volcanic succession (Blake River Group) on the southern side of the fault could not have been deposited north of the fault. Therefore, the fault represented a hinge line south of which the lavas must have undergone considerable subsidence during their deposition. Prehnite, pumpellyite, hydrogarnet, as well as quartz, talc, chlorite, albite, actinolite, epidote and primary olivine, pyroxene, and feldspar occur in rocks of appropriate composition from the base of the Stoughton - Roquemaure Group up to the top of the Blake River Group and in the Destor - Porcupine Complex. Rocks which have been subjected to lower to middle greenschist facies rank metamorphism occur in the Hunter Mine Group.

As the Lake Abitibi Batholith is approached, the metamorphism in the rocks increases from subgreenschist to lower, middle, and upper greenschist facies and at the contact, with the batholith, to amphibolite facies. This is similar to the situation observed by Jolly (1974, 1978) south of Kirkland Lake in the vicinity of the Round Lake Batholith. Metamorphic aureoles of albite - epidote - hornblende, hornblende, and pyroxene hornfels metamorphic rank occur in the volcanic rocks surrounding the alkali intrusions. Alkali metasomatism is common, particularly where the rocks are strongly sheared along the Destor - Porcupine Fault and the Kirkland Lake - Larder Lake Fault Zone.

STRUCTURAL GEOLOGY

REGIONAL LARGE-SCALE STRUCTURAL FEATURES

The largest structural features of the Kirkland Lake area are the synclinorium and the 2 main fault zones, the Destor - Porcupine Fault Zone and the Kirkland Lake - Larder Lake Fault Zone located on the northern and southern limbs of the synclinorium respectively. Jolly (1978) concluded that these 3 structural features formed during the deposition of the volcanic succession, rather than later, and were little affected by the Kenoran Orogeny. Therefore, the large-scale structural development of the area must be considered as having been associated with the origin of the rocks.

The history of the earlier volcanic piles on the margins of the basin, currently represented by the synclinorium, is uncertain. The felsic volcanic rocks of these volcanic piles, as represented by the Hunter Mine and Skead Groups, suggest that they had reached a late stage in their volcanic evolution and had become relatively stable masses of sialic crustal rock. Faults along the margins of these crustal blocks may have been the precursors of the Destor - Porcupine and Kirkland Lake - Larder Lake Fault Zones (*see* Figure 6a).

The deposition of sediments of the Porcupine Group (Pyke 1980) was followed by the formation of a thick succession of ultramafic and mafic lavas of the Stoughton - Roquemaure, Larder Lake, and Kinojevis Groups in the basin, and in low areas on the marginal shelves of the older volcanic piles. As the volcanics accumulated, their weight would produce subsidence in the centre of the basin. This in turn would cause strata around the edges of the basin to be tilted in the directions of the greatest lava accumulations, new fractures to be developed, and older fractures to be reactivated. As areas with thin accumulations of mafic lava remained buoyant and areas with thick accumulation tilted or sank, older rocks would be juxtaposed against younger rocks along faults (*see* Figure 6a).

As the basin, sank, and volcanism represented by the Kinojevis and Blake River Groups became focused in the basin itself, the accommodation of the downward displacement would have taken place along the main marginal fault zones rather than the general sagging envisaged by Goodwin (1977b). The area would comprise the basin acting as a sinking fault block and the older volcanic piles acting as buoyant fault blocks around the edges (*see* Figure 6b). The movement likely occurred along a zone which consisted of a number of faults separated by small fault blocks rather than a single major fault.

The sinking of the basin centre would have caused a tilting inward of the outer volcanic lavas and, thus would have created the synclinorium (*see* Figure 6b). At the same time, this would have produced tensional forces which probably acted along the already active fault zones. The effect of such forces would have been to open the fault zones and form a graben structure in which the Timiskaming Group and the Destor - Porcupine Complex would have been developed and preserved.

The effects of the emplacement of the Round Lake and Lake Abitibi Batholiths are uncertain. These effects probably created domes in the older volcanic piles and possibly some lateral shortening similar to that described by Gorman *et al.* (1978). The only lateral shortening that can be related positively to the batholiths, however, is the development of schistosity and small folds in the volcanic rocks adjacent to the batholiths and possibly, the closure of the grabens.

SMALL-SCALE STRUCTURAL FEATURES

In the Kirkland Lake area, most of the rocks, except those along shear zones in the fault zones and those at the margins of the batholiths, show little or no penetrative deformation. Pillows, amygdules, variolites, rock-fragments, and hyaloclastic textures are undeformed throughout the volcanic sequence. Most of the differential movement in the rocks has occurred along narrow strike-slip fault zones which probably formed during depositional tilting and folding of the strata. Some additional movement could have occurred along these strike-slip faults at a later time.

The absence of penetrative deformation indicates that the attitudes in the volcanic and sedimentary rocks are more closely related to the history of volcanic deposition rather than to post-depositional regional tectonic events. Therefore, movements in the strata such as tilting, folding, and strike-slip faulting will be discussed in conjunction with the deposition of the groups of volcanic rock and emplacement of the Lake Abitibi Batholith.

POST-DEPOSITIONAL FAULTS

Many vertical and near-vertical faults transect the area in various directions. Where exposed, these faults are narrow shear zones which do not offset greatly the volcanic strata or each other, either vertically or horizontally. Most of these appear to be normal tensional faults caused by regional stresses in the Archean crust, formed at various times since the deposition of the volcanic rocks. Diabase dikes have penetrated along some of the north- , northeast- , and northwest-trending faults.

SUMMARY

The volcanic rocks in the Kirkland Lake area are composed of supergroups that have a basal group of komatiitic lavas, upon which are groups of tholeiitic, calc-alkalic, and alkalic lavas, respectively. The formation of these supergroups occurred during cycles of ultramafic - to - felsic volcanism. At the initiation of each cycle, komatiitic lavas as well as some tholeiitic lavas were extruded to form a basal succession. Succeeding the komatiitic volcanism, a succession of iron-rich tholeiitic lavas was deposited followed by a period of calc-alkalic volcanism during which some tholeiitic flows were deposited with the calc-alkalic rocks. Alkalic lavas which overlie the calc-alkalic successions mark the end of each volcanic sequence.

Two complete volcanic sequences and parts of a third are present in the Kirkland Lake area. A repetition of rock types occurs throughout the stratigraphic column. Therefore, the types of these rocks are described in the section on Petrology of the Igneous Rocks and the stratigraphy of these rocks is described in the section on Stratigraphy.

Petrology of the Igneous Rocks

KOMATIITES

RECENT INVESTIGATIONS

Recent investigations pertaining to komatiites suggest that their petrogenesis, field relationships, and ages have wide implications for the evolution of the earth's early crust and mantle. The noncumulate parts of layered flows and the thin nonlayered flows of peridotitic komatiites are generally assumed to closely represent magmas generated in the earth's mantle (Bickle *et al.* 1977; Arndt 1975; and Nesbitt and Sun 1976). Meteorites, common ultramafic rocks, nodules in kimberlites and alkali basalts and basalts in general, indicate that the mantle had a peridotitic composition (Ringwood 1975; White 1967; O'Hara *et al.* 1975; Green 1973a). The major and trace element abundances found in peridotitic komatiites would require 50 to 80% melting of the mantle at depths from about 100 to 900 km and temperatures ranging from about 1650° to 2200°C (Arndt 1977; Nesbitt and Sun 1976; Sun and Nesbitt 1978a; Green *et al.* 1975; Green 1975; McIver and Lenthall 1974). If Archean geothermal gradients were similar to present day geothermal gradients as suggested by England (1979), Ewing (1979), and P.R.A. Wells (1976), magmas for peridotitic komatiites would have to be generated at even greater depths (Nesbitt and Sun 1976). Such depths are well below the low velocity zone in the present day upper mantle and below depths at which basaltic magmas are believed to be generated at the present time (D.H. Green and Ringwood 1967a; Ringwood 1975; Yoder 1976). This suggests that the structure of the Archean mantle was quite different from that of the mantle at the present time (Sun and Hansen 1975a; Green 1975).

Compositions of komatiitic magmas point toward an undepleted mantle with a composition similar to garnet lherzolite (Bickle *et al.* 1977) and to carbonaceous chondrite (Sun and Nesbitt 1977). Arndt (1977) has expressed reservations about calculating the earth's Archean mantle from komatiitic magmas. A rising diapir similar to the model of Ramberg (1972) is generally envisaged to have been involved in the generation of komatiitic magmas and their transportation to the surface (Brooks and Hart 1972; McIver and Lenthall 1974; Cawthorn 1975; Green 1975; Nesbitt and Sun 1976; Bickle *et al.* 1976 and Arndt 1977). Except for Arndt (1977), most of the authors favour a single adiabatic melting episode followed by a sudden upwelling of magma to the surface.

Arndt (1977) maintained that it is difficult to avoid the separation of liquid and residue after 20% melting, and that the peridotitic komatiites of Munro Township display light rare earth element depletions by an earlier melting episode (Arndt *et al.* 1977; Whitford and Arndt 1978). Thus, Arndt's tholeiitic and basaltic komatiite magmas could have been formed by earlier partial melts of a diapir. In contrast, Sun and Nesbitt (1977, 1978a, 1978b) and Nesbitt *et al.* (1979) maintain that peridotitic komatiite magma sources, although heterogeneous, were undepleted to very weakly depleted in the incompatible elements. Nesbitt *et al.* (1979) pointed out that according to their calculations, Arndt's tholeiitic basalts could not have been derived from the same sources as those that produced the komatiites. Instead, the authors suggested that tholeiites, basaltic komatiites, and peridotitic komatiites must have originated from separate sources.

The tectonic environment and significance of komatiites in Archean greenstone belts are open to question (Nesbitt *et al.* 1979). Komatiites tend to be concentrated at the base of greenstone successions (Anhaeusser 1971a, 1971b, 1973; Viljoen and Viljoen 1969b; Glikson 1972, 1976a, 1976b, 1976c, 1976d, 1979a; Arndt *et al.* 1979). Viljoen and Viljoen (1969b) suggested that komatiites must represent Archean primitive crust which was more or less gradational into the primitive mantle. Green (1972a) proposed that greenstone successions are terrestrial equivalents of lunar maria and that komatiitic volcanism was triggered by giant impacts. Shock impact structures, however, are not common in much older rocks (Bickle *et al.* 1977; Glikson 1976d), but Glikson (1976d) did not exclude meteorite impacts as a possible mechanism.

A seafloor environment for komatiite formation has been strongly favoured by several authors (Hallberg and Williams 1972; Herrmann *et al.* 1976; Anhaeusser 1973; Glikson 1976a, 1976b, 1976c, 1976d). McCall and Leishman (1971) described komatiites as being deposited in a eugeosynclinal environment, but Nisbet *et al.* (1977), Nesbitt and Sun (1976), Hawkesworth and O'Nions (1977), have suggested a rift environment. Nisbet and Pearce (1977), proposed that such rifting occurs in a sialic crust and that komatiites of the Belingwe Belt were deposited in a miogeosynclinal environment. Similar environments have been proposed for the formation of komatiites that are exposed in the Northwest Territories, Canada, by Schau (1977) and in Western Australia by Barley *et al.* (1979).

Felsic (calc-alkalic) volcanic rocks are intercalated with the komatiites in many volcanic successions. Calc-alkalic volcanic rocks form major parts of some greenstone belts (Baragar 1968; Jensen 1976a, 1978a). Such observations have led Brooks and Hart (1972, 1974) to suggest an island arc environment for the generation of komatiites, but Nesbitt and Sun (1976), Hawkesworth and O'Nions (1977), and Moores (1973) have strongly argued against this.

Thus, it is apparent that the chemistry and field relationships of komatiites of Archean greenstone belts yield little or no consistent information about the tectonic environment in which these rocks formed.

Proterozoic and Phanerozoic volcanic rocks having similar characteristics to komatiites occur in fold belts (Schwarz and Fujiwara 1977; Peredery 1979), ophiolite assemblages (Gale 1973; Simonian and Gass 1978), island arc environments (Nisbet *et al.* 1977), and rifted continental margins (Clarke 1970). Sun and Nesbitt (1978b), however, have shown low-Ti basalts in island-arc environments, and ophiolite assemblages purported to be komatiites are quite distinct in their chemistry and origin from komatiites. Nevertheless, these authors do accept the Proterozoic Cape Smith volcanic rocks of Schwarz and Fujiwara (1977) as being acceptable komatiites. Arndt *et al.* (1979), believed that Proterozoic and Phanerozoic peridotitic komatiite has yet to be identified. In post-Archean terrains, containing volcanic rocks, only basaltic rocks similar to basaltic komatiites have been found. The Tertiary basalts of Baffin Bay (Clarke 1970) have more than 20% MgO, but are excluded as komatiites because of their high TiO_2 content (Nesbitt and Sun 1976). The Baffin Bay basalts must have required temperatures of melting and depths of origin for their formation similar to those for komatiites (Clarke 1976). This has led Hawkesworth and O'Nions (1977) to propose a rifting environment for komatiites. Baragar and Lamontagne (1980) have reported finding peridotitic komatiites with spinifex olivine in the Ottawa Islands of Hudson Bay. These rocks are considered to be an extension of the Proterozoic Cape Smith fold belt rocks. No chemical data, however, was given.

NOMENCLATURE AND CLASSIFICATION

Komatiitic volcanic rocks were first named and described in the Barberton Mountain area in South Africa by Viljoen and Viljoen (1969b), who argued that many of the ultramafic rocks in the Barberton Mountain Land area were extruded peridotitic lavas because they contained crystalline "quenched" structures of large, bladed olivine crystals later named "spinifex textures" by Nesbitt (1971). Viljoen and Viljoen (1969c) proposed that peridotitic komatiites be distinguished from more familiar ultramafic and mafic rocks by a combination of chemical features composed of the CaO/Al_2O_3 ratio, FeO/MgO ratio, silica content and alkali content. Brooks and Hart (1974) quantified these characteristics and suggested that komatiites were non-cumulative rocks with MgO more than 9%, K_2O less than 0.5%, TiO_2 less than 0.9% and CaO/Al_2O_3 greater than 1.0. Viljoen and Viljoen (1969b) also proposed that peridotitic komatiites have more than 20% MgO by weight and basaltic komatiites have approximately 9 to 20% MgO by weight. In addition, these authors established the Barberton, Badplaas, and Geluk types of basaltic komatiite with average MgO contents of about 10, 15, and 20% by weight, respectively.

Nesbitt and Sun (1976), Nisbet *et al.* (1977), and Arndt *et al.* (1977) pointed out that ultramafic lavas in many greenstone belts share most of the characteristics of the Barberton Mountain Land area peridotitic komatiites, except for the high CaO/Al_2O_3 ratios. Nesbitt and Sun (1976) suggested that the CaO/Al_2O_3 ratio should be about 1 for primary melts, with a probable lower limit of 0.9 for spinifex textured peridotitic komatiite. A CaO/Al_2O_3 ratio of 0.9, however, still excludes many of the lavas of the Abitibi greenstone belt considered to be komatiitic lavas by Arndt *et al.* (1977), and Jensen (1976a, 1976b). Recently, Nesbitt *et al.* (1979), have proposed that peridotitic komatiites can be classified either as aluminum-depleted, or as aluminum-undepleted peridotitic komatiites.

Nesbitt and Sun (1976) also pointed out that Viljoen and Viljoen's (1969b) terms for Geluk, Barberton, and Badplaas types of basaltic komatiite are difficult to apply to other areas where komatiites are exposed. An alternative classification proposed by Arndt (1975) was that komatiitic lavas with 12 to 20% MgO be distinguished as pyroxenitic komatiites and basaltic komatiites should be reserved for lavas from less than 8% to 12% MgO by weight. Implicit in Arndt's classification is that peridotitic, pyroxenitic, and basaltic komatiites form a magma series (Arndt *et al.* 1977). Arndt's classification has been employed in papers by Arndt *et al.* (1977), Naldrett and Arndt (1976), Naldrett and Turner (1977). Other researchers have tended to refer to volcanic rocks with MgO contents between 10 and 20% by weight as either basaltic komatiites (Brooks and Hart 1974; Bickel *et al.* 1975; Anhaeusser *et al.* 1975; Jensen 1976a, 1976b, 1978a, 1978b, 1978c, 1978d, 1978, Nisbet *et al.* 1977; Villaume and Rose 1977) or as high-Mg basalts (Nesbitt 1971; Hallberg and Williams 1972; Hallberg *et al.* 1976; Jolly 1975, 1977).

The nomenclature of komatiites depends on whether or not komatiites represent a separate magma series derived by chemical differentiation from a primary peridotitic magma. Several discriminant diagrams have been developed to distinguish komatiitic rocks from other rock types (Viljoen and Viljoen 1969b; Arndt 1975; Nesbitt and Sun 1976; Jolly 1975; Jensen 1976a). Almost all of these diagrams reveal a 'komatiitic trend' that merges with the start of a tholeiitic trend so that a problem arises as how to distinguish tholeiitic lavas from komatiitic lavas. Francis and Hynes (1979) suggested that komatiites and tholeiites represent a single magmatic suite. On the other hand, Nesbitt *et al.* (1979), have suggested that peridotitic komatiite, basaltic komatiite, and tholeiitic basalts represent distinctive primary magmas.

Arndt (1975), Naldrett and Arndt (1976), Arndt (1976), Arndt *et al.* (1977), and Naldrett and Goodwin (1977) suggested the use of weight percentage Al_2O_3 vs FeO^*/FeO^* + MgO to distinguish the tholeiitic lavas from komatiitic lavas, where peridotitic komatiites are associated with less mafic lava flows. Green and Schulz (1977) pointed out that the iron-rich basaltic komatiites from the Vermillion district of Minnesota, as well as the average Badplass and the average Barberton basaltic komatiite of Viljoen and Viljoen (1969a), were identified as tholeiites when using the discriminant diagram of Naldrett and Arndt (1976). Nesbitt *et al.* (1979) also found that high MgO - low Fe types of mid-ocean ridge basaltic glasses could be classified as komatiites.

Nesbitt *et al.* (1979) pointed out that the rocks classified as tholeiitic basalts by Arndt *et al.* (1977), in Munro Township must have had parental liquids which would be basaltic komatiites according to Arndt (1975). Thus, many of the volcanic rocks with less than 10% MgO described as basaltic komatiites by Arndt *et al.* (1977), could be tholeiitic volcanic rocks. This could be the case where these rocks are described as having "textures and modal compositions not unlike those of non-komatiitic basaltic rocks" (Arndt *et al.* 1977, p.346). The study made by Arndt (1975) on 2 thick differentiated lava flows in Munro Township, Ontario, strongly suggests that rocks of tholeiitic composition can form from liquids of komatiitic composition, and further points to the probability that rocks with tholeiitic compositions are a component of komatiitic differentiation trends (Jensen 1976a; Francis and Hynes 1979).

Komatiitic differentiation trends represent a magnesium depletion with a corresponding enrichment of aluminum (Jensen 1976a, 1976b). Iron enrichment does not occur in the residual liquids relative to the

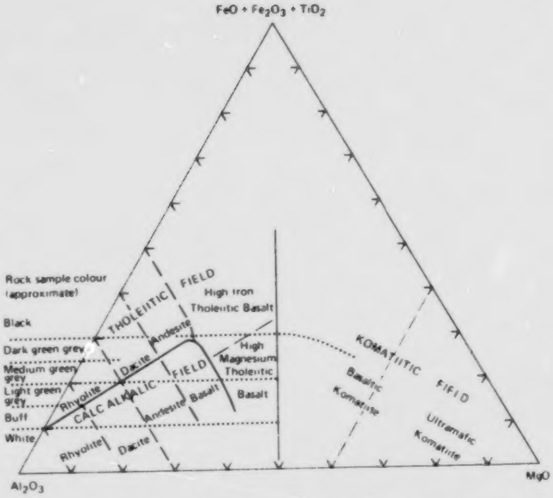


Figure 7. Jensen Cation Diagram involving the cation percentages of Al_2O_3, FeO + Fe_2O_3+ TiO_2, and MgO.

primary liquids, regardless of the iron content in the original liquid, and would not be anticipated upon fractional crystallization of olivine (Donaldson 1975). This is illustrated by analyses from Theo's and Fred's flows that have been studied by Arndt (1975). The high Fe-magma represented by Theo's flow cannot be explained except as a lower degree of partial melting of mantle material. A primary iron-rich magma has been proposed by Pearce and Birkett (1974).

Although Nesbitt *et al.* (1979) concluded that peridotitic komatiites can be recognized and defined by the presence of (olivine) spinifex textures, there is as yet no suitable definition of basaltic komatiites except for the fact that they are high Mg basaltic liquids. Nesbitt *et al.* (1979) has shown that discriminant diagrams are unreliable for defining komatiites. Olivine spinifex, however, is related to the cooling history of a lava flow (Donaldson 1975, 1976, 1977). Under proper conditions, olivine spinifex can form as a texture from a variety of olivine normative liquids. Many peridotitic komatiite lavas do not contain spinifex (Arndt *et al.* 1977), and olivine spinifex has been observed in lava flows with less than 20% MgO (Echeverria, worker at Carnegie Institute, Washington, personal communication, 1979). The presence of olivine spinifex may not be the only criterion for defining a peridotitic komatiite.

A more suitable definition for a peridotitic komatiite would be that it is a volcanic flow with more than 20% MgO. This conclusion is similar to the one reached at the Penrose Conference on komatiites (Arndt and Brooks 1980).

In this study, the term 'basaltic komatiite' is reserved for nonlayered lava flows and the noncumulate portions of thin layered lava flows with 10 to 20% by weight (anhydrous) MgO. This criterion effectively separates the olivine-rich lavas from the pyroxene-rich lavas (Arndt 1975) and it effectively separates the komatiites from the tholeiitic basalts which generally have less than 10% by weight MgO.

Thick layered flows (for example, Theo's Flow and Fred's Flow, see Arndt 1975) and layered bodies which contain peridotitic, pyroxenite, and gabbro, are described individually where these bodies occur in the komatiitic succession. These bodies are restricted to volcanic sequences which contain komatiites, and are considered to be a part of the komatiitic succession.

In this study, the Jensen Cation Diagram (Figure 7) employing the cation percentages of Al_2O_3, Fe_2O_3 + FeO + TiO_2, and MgO on a ternary plot are used to distinguish komatiitic volcanic rocks from other subalkalic volcanic rocks (Jensen 1976a). On this diagram, the boundary between the komatiitic and tholeiitic fields corresponds to MgO = 10% by weight. The line dividing the komatiites into basaltic and peridotitic komatiites is based on petrographic differences. Olivine is the major mafic mineral in peridotitic komatiites, and pyroxene is the major mineral in basaltic komatiites. Komatiites with 20% by weight MgO are projected near or on the line separating basaltic from peridotitic komatiites. Plots of analyses from volcanic sequences containing komatiitic rocks are shown in Figures 8 and 9.

The tholeiitic basalts are divided into Mg-rich and Fe-rich tholeiitic basalts. Their nomenclature and classification are discussed elsewhere in this report. Magnesium-rich tholeiitic basalts are abundant in volcanic sequences that contain komatiitic volcanic rocks, suggesting that these are related to each other. Iron-rich tholeiitic basalts tend to be rare or absent in most komatiitic sequences (for example, Anhaeusser 1973) and where found, their FeO total and TiO_2 contents are less than those in sequences where komatiites are absent (Jensen 1976a).

FIELD AND PETROGRAPHIC CHARACTERISTICS OF KOMATIITES AND THEIR ASSOCIATED INTRUSIVE ROCKS

Much effort has gone into documenting the field and petrographic characteristics of komatiites (Viljoen and Viljoen 1969b, 1969c; Nesbitt 1971; McCall and Leishman 1971; McCall 1971; Williams 1972; Pyke *et al.* 1973; Imreh 1974, 1976; Fleet and MacRae 1975; Arndt 1975, 1976; Arndt *et al.* 1977; Gelinas *et al.* 1977a; Lajoie and Gelinas 1978; Nisbet *et al.* 1977; Arndt *et al.* 1979). Most of these studies have been based on a few exceptionally well exposed outcrops where the primary mineralogy and textures are preserved. Elsewhere, komatiites tend to be poorly preserved because of their susceptibility to alteration and weathering, and therefore, they tend to be difficult to study in the context of their original textures, mineralogy and chemistry.

Different terms have been applied to some field and petrographical features of komatiites. Some terms have come to be generally accepted while others have not. 'Spinifex' texture (Nesbitt 1971), a texture that results from the parallel growth of olivine grains into large, closely spaced, plate-like crystals to form booklets of sheafs, has become a generally accepted term by field geologists. This same texture also has been referred to as "chicken track", "herring-bone", "lath", "feather", or "quench", throughout northeastern On-

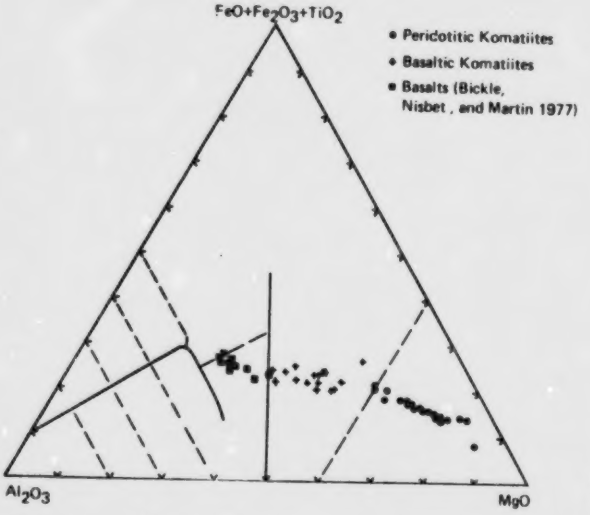


Figure 8. Volcanic rocks of the Belingwe Greenstone Belt, Rhodesia, projected on the Jensen Cation Diagram. Plots according to Bickle, Nisbet, and Martin (1977).

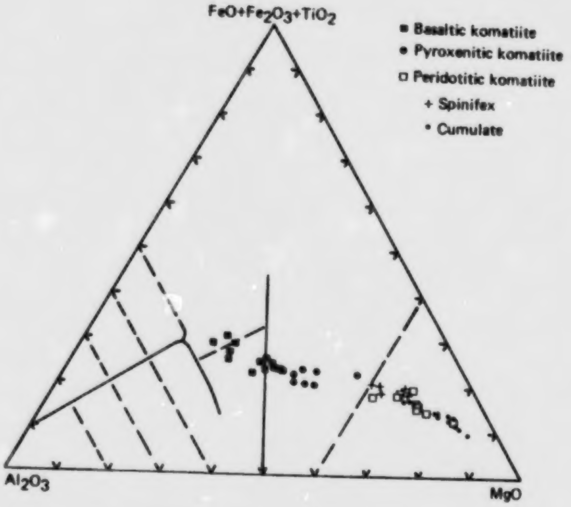


Figure 9. Volcanic rocks of Munro Township, Ontario (Arndt 1975), projected on the Jensen Cation Plot. Rock names are according to Arndt (1975).

tario (Pyke *et al.* 1973). Among experimental petrologists who have studied the various crystal forms of olivine, the term 'plate olivine' has been applied instead of 'spinifex' (Donaldson 1976; Lofgren 1979). 'Spinifex' will probably continue to be used as a general term in the field and that 'plate olivine' will be used for a very specific type of olivine crystal.

The plate-like form of olivine, called 'spinifex' texture by Nesbitt (1971), and the 'string-beef' clinopyroxene spinifex of Arndt (1975), are only 2 of several crystal types of olivine and pyroxene that can be recognized in well preserved komatiitic lavas (Arndt *et al.* 1977; Nisbet *et al.* 1977). Experimental studies by Lofgren *et al.* (1975), Usselman *et al.* (1975), Donaldson (1974, 1975, 1976, 1977), and Donaldson *et al.* (1975) have achieved considerable success in reproducing the variety of crystal types of pyroxene and olivine found in komatiites. In their experiments, these authors were able to show that a hierarchy of olivine and pyroxene crystal types and fabrics exists. This hierarchy is related to the rate of cooling of magmas under various conditions involving initial compositions and temperatures, degrees of initial crystal nucleation, and oxygen fugacity (Lofgren 1979). From this hierarchy, much can be said about the cooling history of a komatiitic lava flow by studying the crystal shapes and textures of olivine and pyroxene in each part of the flow. Such a study, however, is beyond the scope of this report which is confined to the field and petrographic characteristics of komatiites that pertain to the interpretation of the Kirkland Lake area.

As yet, there is little consistency in the terminology to describe the pattern of thin joints or fractures separating rounded to polygonal-shaped blocks or fragments of variable size at the top of and within komatiitic lava flows. For example, Arndt (1975, 1976), Arndt *et al.* (1977, 1979), Pyke *et al.* (1973), Pyke (1970, 1975, 1978a, 1978b, 1980) use the terms "polygonal jointing", "polygonite", "polyhedral jointing", "polygonal fracturing", and "polysuturing" synonymously. In this report, the terminology used by Arndt *et al.* (1977) to describe this "fracturing pattern" in komatiitic lava flows will be adhered to as closely as possible.

Peridotitic Komatiites

Peridotitic komatiites form well-rounded, poorly exposed outcrops having soft, rusty brown weathering surfaces 0.5 to 1.5 cm thick. When fresh, the rock is dark bluish grey to greenish black depending on the degree of serpentinization.

Most peridotitic komatiites occur as tabular flows ranging from 1 to 5 m in thickness, or as pillows 0.1 to 1.0 m across. Some flows can be made up of lava toes or lava lobes (Arndt 1975). The tabular flows range from being nonlayered to strongly layered. The layering is caused by varying amounts of subequant to equant olivine crystals settling toward the base of the flow, and by skeletal to bladed olivine crystals (olivine spinifex) growing downward from the chilled flow top (Figure 10; Pyke *et al.* 1973; Arndt 1977; Arndt *et al.* 1977). The spinifex olivine is considered to be the result of olivine crystals growing rapidly in a magma having a high normative olivine content during supercooling (Donaldson 1976, 1977; Walker *et al.* 1976). This probably occurred in the magma shortly after it erupted; the olivine phenocrysts had a chance to settle to the base of the flow when it had come to rest and still had a low viscosity (Donaldson 1974; Arndt 1976; Arndt *et al.* 1977). Where found, such flows are useful for top determinations because the relict textures of spinifex olivine tend to be preserved in spite of extensive alteration of the olivine to antigorite, talc, magnesite, and magnetite. Below the chilled and fractured flow top, the olivine spinifex tends to become

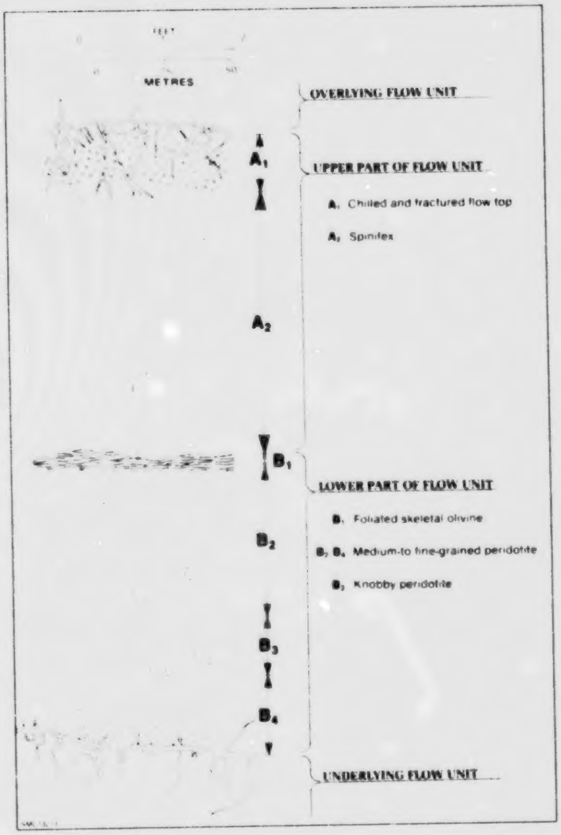


Figure 10. Diagrammatic section of a typical ultramafic flow in Munro Township (after Pyke, Naldrett, and Eckstrand 1973).

more elongate and perpendicular to the strike of the flow (Figure 10).

The crystallization of olivine caused crystal-liquid fractionation in the magma which resulted in skeletal to acicular pyroxene and chromite crystals forming interstitially among the cumulate olivine grains and spinifex olivine grains. In thick flows, such as those described by Arndt (1976), slower cooling enabled cumulate layers of pyroxene to form above the olivine cumulates (Figure 11). Toward the top of the flow, skeletal pyroxene formed below the skeletal olivine. The remaining liquid crystallized as a normal gabbroic rock.

Arndt *et al.* (1977) suggested that layered flows with cumulate olivine and spinifex olivine were formed from magmas of low viscosity. Such magmas, as lavas, would flow rapidly and tend to be thin, unless they were dammed by obstructions or ponded in depressions. The effect of such flows would be to smooth out the surface on which they were being extruded. If so, thick flows could be expected toward the base of a komatiitic succession and thinner and more widespread flows could be expected higher in the succession. Some evidence of this is found in the Kirkland Lake area.

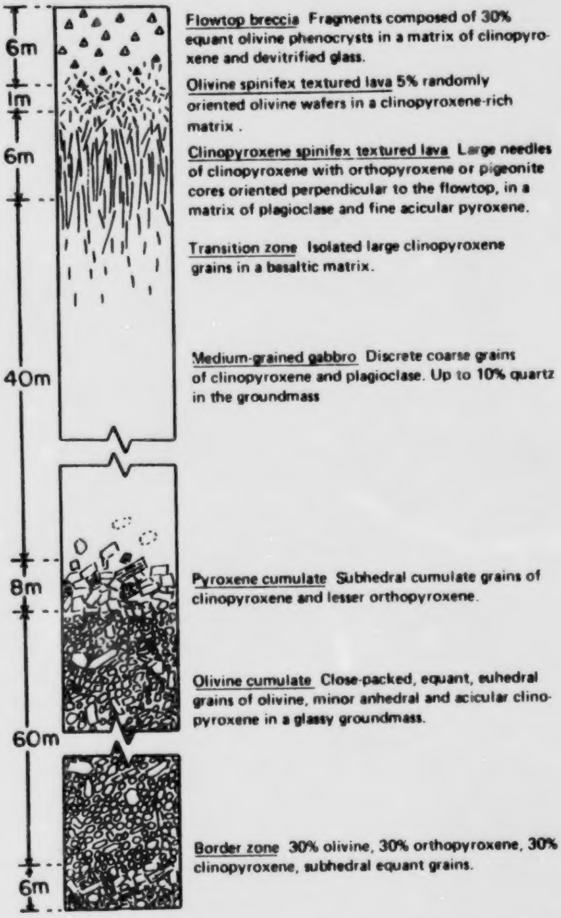

6m
1m
6m
40m
8m
60m
6m
Flowtop breccia Fragments composed of 30% equant olivine phenocrysts in a matrix of clinopyroxene and devitrified glass.
Olivine spinifex textured lava 5% randomly oriented olivine wafers in a clinopyroxene-rich matrix.
Clinopyroxene spinifex textured lava Large needles of clinopyroxene with orthopyroxene or pigeonite cores oriented perpendicular to the flowtop, in a matrix of plagioclase and fine acicular pyroxene.
Transition zone Isolated large clinopyroxene grains in a basaltic matrix.
Medium-grained gabbro Discrete coarse grains of clinopyroxene and plagioclase. Up to 10% quartz in the groundmass
Pyroxene cumulate Subhedral cumulate grains of clinopyroxene and lesser orthopyroxene.
Olivine cumulate Close-packed, equant, euhedral grains of olivine, minor anhedral and acicular clinopyroxene in a glassy groundmass.
Border zone 30% olivine, 30% orthopyroxene, 30% clinopyroxene, subhedral equant grains.

Figure 11. Diagrammatic section through Fred's Flow - a komatiitic layered flow (after Arndt 1975).

Many flows of peridotitic komatiite are massive and have little or no spinifex olivine. Spinifex olivine is believed to develop only in magmas essentially devoid of olivine phenocrysts (Donaldson 1974). Arndt *et al.* (1977) concluded that flows without spinifex olivine were caused by lavas becoming sufficiently viscous to retard or prevent the settling of olivene phenocrysts. Hence, these flows are essentially homogeneous and featureless except for their pervasive polyhedral jointing (Figure 12) (Arndt *et al.* 1977). The joints are close spaced and the fragments subangular near the tops of the flows. Downward in the flows, the polyhedral joints become spaced more widely and the polyhedra larger. A thin, chilled margin which is distinctly more deeply weathered and wider than the polyhedral joints, occurs at the tops of the massive flows. In places, a thin layer, 0.5 to 3 cm in thickness composed of hyaloclastite, occurs above the chilled margin.

The origin of the polyhedral jointing is uncertain (Arndt 1975). It appears to represent fractures developed as a result of contraction during cooling, when the flows were nearly solidified. Liquid magma inter-

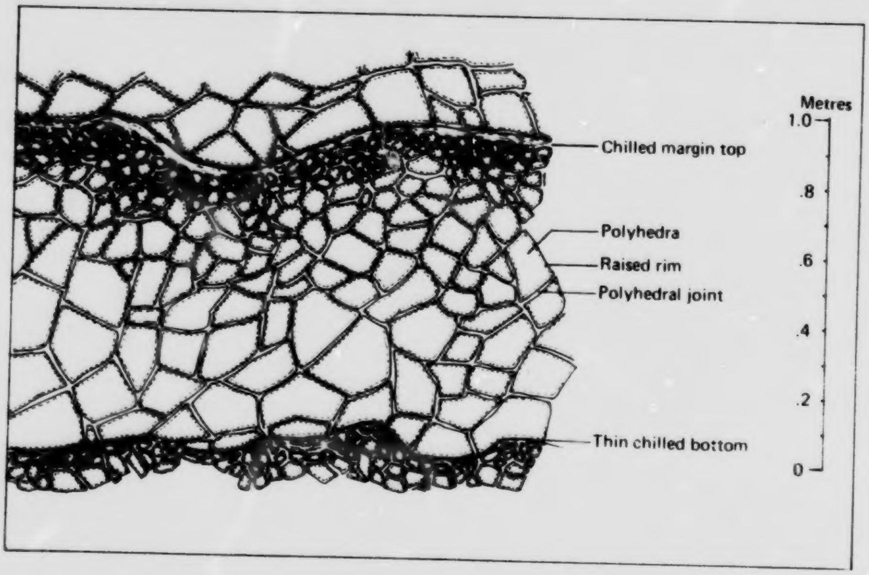


Figure 12. Section diagram of a tabular nonspinifex peridotitic komatiite flow.

stitial to the olivine grains appears to have seeped into the polyhedral joints of some flows and caused additional olivine growth on the surfaces of the polyhedra (Arndt 1975). Some polyhedral joints are filled with tremolite, suggesting that acicular clinopyroxene also developed in the joints.

Pillowed peridotitic komatiitic lavas are rare, but where seen, pillows have black to blue-black, serpentine-rich chilled margins, 1 to 2 cm thick, which are soft and more deeply weathered than the cores of the pillows. Pillow tails tend to be shorter and smaller than those of other types of lava; and the crevices between the pillows are filled with carbonate and serpentinized volcanic fragments. The pillow cores are polyhedrally jointed. Individual polyhedra, 2 to 5 cm in size, are similar to those found toward the top of massive peridotitic komatiitic flows. The pillowed lavas are very similar to those described by Nisbett *et al.* (1977).

Relict polyhedral jointing and spinifex textures are preserved even in highly metasomatized and metamorphosed flows, provided there has been no tectonic deformation. In environments of tectonic stress, the polyhedra become rounded and elongated parallel to the shearing, and, as with the spinifex textures, are ultimately destroyed. Peridotitic komatiites metamorphosed to assemblages of talc, actinolite, tremolite, chlorite, and serpentine-related minerals have a lighter colour and have a clotty fibrous texture in which polygonal fragments and spinifex textures remain easily identifiable by the naked eye.

Near the contact with the Lake Abitibi Batholith, the peridotitic komatiites are hard, dark green, schistose amphibolites in which the polyhedra are either strongly elongated or destroyed. Away from the batholith, the peridotitic komatiites become lighter coloured and softer; the polyhedral jointing is retained. Under conditions of intense shearing and low temperatures along the Destor - Porcupine Fault, ultramafic komatiites become soft, grey to light grey talc and talc-chlorite schists having little or no chemical change from that of the original rock except for the addition of H_2O.

Peridotitic komatiites are very susceptible to changes in composition and appearance in zones of hydrothermal alteration, particularly near intrusions of alkalic rock. A grey, brown, orange or emerald-green massive to schistose rock can result, which is called 'carbonate' in the Kirkland Lake area (Thomson 1950). This type of rock formed from peridotitic komatiite by the introduction into it of hydrous fluids and gases carrying carbon dioxide and alkalies. Fine-grained ferruginous magnesite and dolomite form 10 to 25% of the 'carbonate' rock. Various amounts of sericite, chlorite, talc, anthophyllite, and quartz form the remainder of the rock. Quartz derived from the conversion of olivine to magnesite can form veins in the rock. Chrome mica gives the 'green carbonate' rock its emerald green colour.

Basaltic Komatiites

Basaltic komatiites form low, flat, well rounded outcrops having soft brown to greyish brown weathering surfaces 0.5 to 1.0 cm thick. The rocks are soft and crumbly, and greenish grey when fresh. As a group, basaltic komatiites tend to be quite variable in mineralogy and texture because of their wide range of chemical compositions. Textures within a flow are greatly influenced by the cooling history of the flow (Arndt *et al.* 1977).

Basaltic komatiites occur as tabular flows 3 to 20 m thick and as pillowed lava units 2 to 30 m thick. The tabular flows and pillowed lava horizons have strike lengths of several hundred m in areas of good outcrop. In places, the tabular flows can grade either laterally or vertically into pillowed lava. The tabular flows can either be massive and homogeneous, or have layers of pyroxene and olivine with crystal forms of dendritic, skeletal, and euhedral shape.

TABLE 1: OUTLINE OF ALTERATION SEQUENCE.

	→ Increasing alteration →		
Olivine	→ serpentine and magnetite, chlorite in less ultramafic komatiites	→ serpentine, magnetite tremolite, talc, and/or chlorite	→ random aggregate of serpentine, tremolite, talc chlorite, magnetite
Clinopyroxene	→ clinopyroxene (except fine plumose sheaves replaced by tremolite)	→ tremolite and chlorite	→ random aggregate of serpentine, tremolite, talc chlorite, magnetite
Groundmass	→ Devitrified glass too fine grained to resolve optically	→ chlorite, tremolite, serpentine, talc, and ore	→ random aggregate of serpentine, tremolite, talc chlorite, magnetite
	Igneous textures visible		No igneous textures visible

(After Nisbet *et al.*1977)

The sequence of alteration of basaltic komatiite is similar to that described by Nisbet *et al.* (1977) (Table 1). In some basaltic komatiites of the Kirkland Lake area, the olivine, pyroxene, and plagioclase components are sufficiently preserved that description of the primary textures is possible.

Among komatiitic lavas, olivine decreases from 35 to less than 1% as their MgO content decreases from 20 to 12% (Arndt *et al.* 1977). Olivine-rich basaltic komatiites tend to be massive, featureless and granular, except for their aphanitic flow tops which may be brecciated and contain small (1 to 3 mm) skeletal randomly oriented wafer-shaped (plate) olivine grains. Such flows tend to be polyhedrally jointed as shown in Figure 12. In the flow tops, olivine occurs as equant phenocrysts with skeletal overgrowths, as discrete prismatic grains, and as very thin, plate-like skeletal grains which are set in a matrix of acicular dentritic or plumose clinopyroxene, sparse euhedral grains of chromite and devitrified glass, as described by Arndt *et al.* (1977). In the remainder of the flow, the olivine grains are equant to subequant (0.5 to 1.5 mm) and are separated by smaller, subequant and skeletal clinopyroxene grains.

In some olivine-rich flows of basaltic komatiites, plate olivine (Donaldson 1976) can extend down from the top to give the rock a hackly surface for more than half of the flow's thickness. The plates of olivine may increase from 1 to 3 mm in size at the top of the flow to plates 2 to 3 cm in size at a depth of 1 to 2 m below the top. Below this level olivine plates may decrease in size and grade into the more granular lower part of the flow. Some flows have 2 or more zones of large olivine plates alternating with zones of small plate olivine. The lower parts of such flows have 30 to 60% equant olivine grains. This suggests that accumulation of olivine has occurred.

Olivine-rich basaltic komatiitic flows with lower MgO contents, have zones (3 to 50 cm thick) of closely packed, needle-shaped skeletal pyroxenes (1 to 20 cm long) interlayered with, and below the zones of plate olivine. The pyroxene grains are perpendicular to the top of the flow, as observed by Arndt (1976).

Pyroxene-rich basaltic komatiitic flows tend to be massive and featureless and "tremolitic". Polyhedral jointing is either weak or lacking except at the top of the flows. The flow tops contain either very fine grained pseudomorphs of randomly oriented plate olivine or acicular pyroxene set in a matrix of spherulitic pyroxene and devitrified glass. This texture gives way, downward, to randomly oriented elongate to subequant clinopyroxene in the massive flows. In flows of 10 to 12% MgO, fasciculate (Walker *et al.* 1978) intergrowths of pyroxene and plagioclase occur; these give way, downward, to elongate and subequant clinopyroxene with interstitial subhedral plagioclase.

Zones of different pyroxene morphology occur in some pyroxene-rich flows where small randomly oriented, acicular pyroxene can become larger (1 to 3 cm long) downward in the flow, and impart a distinctive 'chicken track' texture to the rock surface. The grains of acicular pyroxene occur in a matrix of spherulitic pyroxene with or without plagioclase and devitrified glass. In other flows, the acicular pyroxene gives way downward to closely packed needle-shape skeletal pyroxene 1 to 20 cm long. Oriented pyroxenes give way downward, either to another zone of randomly oriented acicular pyroxene, or to more granular pyroxene that forms the lower part of the flow. The lower parts of such flows tend to have elongated to tabular clinopyroxene and orthopyroxene which may have cumulated toward the base of the flow.

Pillowed lavas of basaltic komatiites are common and are similar to those described by Nisbet *et al.* (1977), and Arndt *et al.* (1979). The pillows are 0.3 to 1.5 m across and have 0.5 to 2 cm thick, chilled margins. Like the tabular flows, the pillows vary in their textures and appearances, according to the proportion of olivine and pyroxene present. The olivine-rich pillowed lavas have polyhedral jointing as shown in Figure 13 and Figure 1C of Arndt *et al.* (1979), whereas the pyroxene-rich pillowed lavas have less distinct or no polyhedral jointing. Textures found in and adjacent to the margins of the pillows, particularly toward their tops, are the same as those found at and adjacent to the flow tops of tabular flows with similar proportions of olivine and pyroxene. Variolites 1 to 10 mm in size, composed of sperulitic

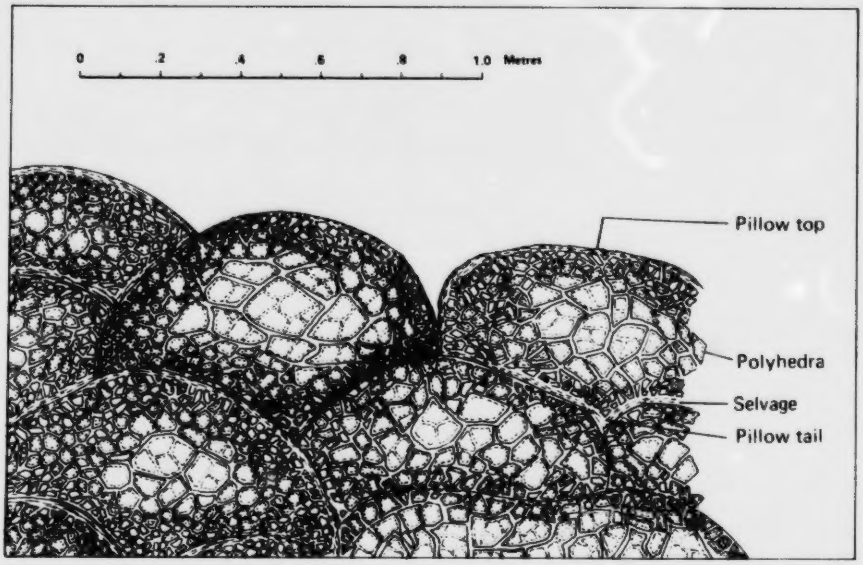


Figure 13. Section, diagram of pillowed basaltic komatiite lavas.

plagioclase and pyroxene, occur in and near the margins of some pyroxene-rich pillowed lavas.

Peridotite Stocks

Elongate to circular stocks of peridotite, 50 to 1500 m across, cut the komatiitic lavas and their underlying rocks. Peridotite stocks do not cut rocks younger than the komatiites, suggesting that the 2 are closely related. The peridotite stocks may have been feeders for some komatiitic flows.

The peridotites are dark bluish grey to green magnetic rocks which weather quickly to white and later, to rusty brown. Numerous joints 1 to 3 mm wide, filled with magnetite, carbonate, talc, and fibrous asbestos occur throughout the rocks. The peridotites consist of 40 to 65% oval antigorite pseudomorphs, that are 0.1 to 5 mm in diameter after olivine. Bastite, tremolite, chlorite, and calcite are pseudomorphs of pyroxene and minor amounts of plagioclase. These pseudomorphs occur between the antigorite pseudomorphs.

ARCHEAN THOLEIITIC VOLCANIC ROCKS

RECENT STUDIES

Archean tholeiitic volcanic rocks have received considerable attention because they are interlayered with komatiites. Much of this work has been centred on the geochemistry and petrography of selected samples of tholeiitic basalt from various greenstone belts. The purpose of this work has been to compare these rocks with Cenozoic tholeiitic lavas; or to establish their petrogenetic associations with other igneous rocks. Geochemical surveys, such as those of Wilson *et al.* (1965), Goodwin (1965), Baragar (1966, 1968), Viljoen and Viljoen (1969c), Glikson (1972), Hallberg (1972), White *et al.* (1971), have been used to establish the fact that Archean tholeiitic volcanic rocks were mainly low-K tholeiites (Arth 1976: Hart *et al.* 1970; Irvine and Baragar 1971, Figure 8D, p.539; Brooks and Hart 1974). Low-K tholeiites occur in almost every modern tectonic environment and show little or no chemical differences from one tectonic environment to another (Brooks and Hart 1974). Therefore, although the presence of Archean tholeiites cannot suggest a unique tectonic environment for greenstone belts, it can be used to reveal considerable information about the development of such belts.

Alkalic rocks appear to be absent from most Archean greenstone belts (Hallberg and Williams 1972; Hawkesworth and O'Nions 1977). This absence has led some to favour either an island-arc or an ocean floor environment for the deposition of Archean tholeiitic rocks rather than an ocean-island setting such as Hawaii. An ocean-floor environment has been preferred by Hallberg and Williams (1972), Glikson (1972, 1976d), Anhaeusser (1973), and Pearce *et al.* (1975), whereas Hart *et al.* (1970), White *et al.* (1971), and Windley (1979), among others have supported the theory of an island-arc environment.

Experimental studies by Green and Ringwood (1967), Nicholls and Ringwood (1972, 1973), and Green (1973b), suggested that a spectrum of tholeiitic magmas could be produced by 25 to 30% melting of pyrolite at depths of 30 to 70 km and temperatures of about 1100 to 1200°C, depending on H_2O content. In contrast, komatiitic magmas appear to require much greater depths and temperatures and higher degrees of partial melting. This suggests that tholeiitic magmas are generated from sources other than that of komatiites, as proposed by Sun *et al.* (1979).

Many authors state that tholeiites with 8 to 9% MgO are primary magmas which have reached the surface with little or no fractionation (Frey *et al.* 1974; Langmuir *et al.* 1978; Maaloe 1979; Sun *et al.* 1979). On the other hand, O'Hara (1965, 1968a, 1968b), O'Hara *et al.* (1975), Yoder (1976), and Elton (1979) have argued that all tholeiitic basalts represent strongly fractionated magmas derived from more magnesian parental melts similar in composition to tholeiitic picrite. This view has also been recently accepted by Green *et al.* (1979), but these authors continued to maintain that tholeiitic picrite was formed at depths not exceeding 70 km at temperatures of about 1430°C.

The significant conclusions drawn from this experimental work are that such tholeiitic magmas fractionate olivine, pyroxene, and plagioclase at depths less than 15 km to produce iron-enriched quartz tholeiite magmas (D.H. Green and Ringwood 1967a), and that tholeiitic magmas cannot fractionate to produce calc-alkalic basalts and associated rocks. Instead, tholeiitic magmas can fractionate only to produce tholeiitic andesite and iron-rich granophyre (T.H. Green and Ringwood 1968), and plagiogranite (Dixon and Rutherford 1979). This strongly suggests that low-K Archean tholeiitic basalts are a suite of rocks distinctive from both komatiites and calc-alkalic volcanic rocks.

Cenozoic low-K tholeiitic basalts tend to have 10 to 20 times chondritic REE values and tend to show LREE depletion, which was probably caused by an earlier partial melting (Gast 1968; Frey *et al.* 1974; Kay and Hubbard 1978; White and Schilling 1978; Sun and Nesbitt 1977, 1978a, 1978b). Unlike their recent counterparts, Archean low-K tholeiites are not depleted in LREE. This shows that their mantle sources did not undergo an earlier episode of melting (Hawkesworth and O'Nions 1977; Fryer and Jenner 1978; Arth and Hanson 1975; Sun and Nesbitt 1978a). Instead, the Archean mantle was essentially chondritic in composition (Zindler *et al.* 1978). In the Kirkland Lake area, flat to slightly LREE-enriched patterns have been obtained in the tholeiites of Munro Township by Condie and Baragar (1974), Arth *et al.* (1977). However, Jackson and Smith (1978) and Jackson (1980) obtained flat to slightly LREE depleted patterns ranging from 8 to 40 times chondrite among the tholeiitic lavas in the Kinojevis Group (*see* Figure 5). In these results, total REE abundances increased with increasing FeO_T/MgO and are similar to the results of Schilling *et al.* (1976) along the Galapagos spreading centre. Here, these authors observed that the REE abundances increased from about 10 to 50 times chondrite as FeO_T content increased from 9 to 18% in low-K tholeiite.

Gill and Bridgwater (1979), in their study of the Ameralik dikes, observed that these dikes of low-K tholeiites resembled Archean greenstone belt tholeiites except for their marked iron enrichment. These authors suggested that basalts with about 16% FeO_T, which they called ferro-basalts, did not occur commonly in greenstone belts. This implies that the Kinojevis Group represents a group of tholeiitic volcanic rocks not normally associated with greenstone belts, although basalts with high FeO_T contents do exist elsewhere (Williams 1972; Naldrett and Turner 1977). The range of rock types in the Kinojevis Group is very similar to those of the Galapagos spreading ridge reported by Byerly *et al.* (1976).

Byerly *et al.* (1976), Schilling *et al.* (1976), and Gill and Bridgwater (1979), concluded that their iron-rich tholeiitic basalts were derived by the fractionation of olivine, plagioclase, and pyroxene from normal tholeiitic magmas at shallow depths.

NOMENCLATURE AND CLASSIFICATION

The term 'tholeiite' was first used in 1840 by Steininger for naming basaltic rocks near Tholey, Saarland, Germany (American Geological Institute Glossary 1970). However, the term 'tholeiite' did not receive prominence until Kennedy (1933) and Tilley (1950) applied the term to a magma type. Reviews of this historical development concerning the use of 'tholeiite' to refer to a saturated to oversaturated magma series showing an iron enrichment trend, have been given by Turner and Verhoogen (1960), Barth (1962), Wilkinson (1967), and Irvine and Baragar (1971).

Since 1970, many major element classifications of volcanic rocks have been proposed to distinguish the tholeiitic rocks from the alkaline volcanic and the calc-alkalic volcanic rocks, and also to distinguish tholeiitic rocks from different tectonic environments (Irvine and Baragar 1971; Middlemost 1972, 1974, 1975; Pearce 1974; Miyashiro 1974, 1975; Jolly 1975; Church 1975; Le Maitra 1976, Jensen 1976a; Delong and Hoffman 1975; Johnson 1979). Examinations of all these classifications are beyond the scope of this study. Except for Jensen (1976a) none of the classifications distinguishes more than one type of tholeiitic basalt.

Tholeiitic basalts form more than 90% of the rocks found in tholeiitic suites. Tholeiitic picrite basalts, and tholeiitic andesites, dacites and rhyolites are uncommon. This suggests that most of the crystal-liquid fractionation of tholeiitic magmas occurs within the compositional range represented by tholeiitic basalt. Thus, Gelinas *et al.* (1977a, 1977b), have described the tholeiitic successions of the Noranda area as monotonous sequences of metabasalt. In order to discern variations better within these basaltic sequences, Jensen (1976a) has proposed a classification which further divides the tholeiitic basalts into 2 or more rock types recognizable in the field (*see* Figure 7).

In several townships north of Kirkland Lake, basalt classified only as tholeiitic basalt by Irvine and Baragar (1971), was divided into 2 types by Jensen (1973a, 1973b, 1973c). The first was a grey basalt having generally more than 6% MgO and less than 12% FeO_T by weight; the second was a black, magnetic, basalt having generally less than 6% MgO and more than 12% FeO_T by weight. Jensen (1976a) proposed that the 2 types be named magnesium-rich and iron-rich tholeiitic basalt (Figures 7, 14).

Because no other studies of this type have split the tholeiitic rocks into rock types, representative chemical analyses of the various tholeiitic volcanic rock types used in this study are shown in Tables 2 and 3.

Pearce and Birkett (1974) made a study on these 2 types of basalts and proposed that the iron-rich tholeiitic basalt represented a distinctive magma type. These authors considered the magnesium-rich tholeiitic basalt to resemble a typical tholeiite. Jensen (1973a) noted further that tholeiitic andesite, dacite, and rhyolite were associated with both types of tholeiitic basalt, which suggests that all these rock types probably were formed by fractionating plagioclase and pyroxene from a magma similar in

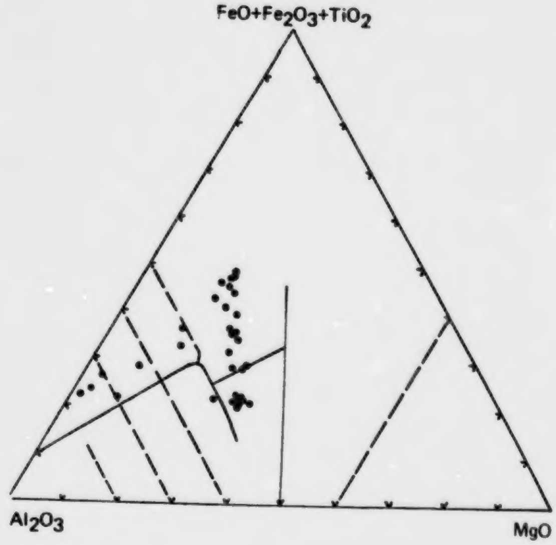


Figure 14. Tholeiitic volcanic rocks of the Galapagos Ridge projected on the Jensen Cation Diagram. Analyses from Swanson *et al.* (1974), Byerly *et al.* (1976), and Schilling (1976).

composition to magnesium-rich tholeiitic basalt (Jensen 1978a; Jackson 1980).

Naldrett and Goodwin (1977) suggested that the weight percentage Al_2O_3 vs $FeO_T(FeO_T + MgO)$ be used to distinguish iron-rich tholeiite from other rock types. These authors used this classification, and noted that iron-rich tholeiites occurred in Garrison Township and suggested that the remaining low-iron rocks with less than 16.5% Al_2O_3 in the area be considered as komatiites. The problems associated with the use of this classification have been discussed elsewhere.

FIELD AND PETROGRAPHIC CHARACTERISTICS

Detailed descriptions of the tholeiitic volcanic rock types in parts of the Kirkland Lake area have been made by Jensen (1978a, 1978b) and Jackson (1980). In this report, these descriptions will be summarized in terms of the experimental work on rock textures by Lofgren (1979). Much has yet to be discerned about the cooling history of tholeiitic lava flows. A start on this type of research has been made on the tholeiitic basalts in the Noranda area by Gelinas and Brooks (1974), Gelinas *et al.* (1977a, 1977b), Dimroth and Lichtblau (1979) and, in Thackeray Township by Pearce and Birkett (1974). Similar research has been carried out on submarine basalts (for example, Bryan 1972; Langmuir *et al.* 1977; Rhodes *et al.* 1979). In the future, it seems as if much of the description of tholeiitic lavas will be done in terms of the mineral morphologies within them in order to interpret their cooling histories and modes of extrusion.

Intrusive phases of the tholeiitic lavas are rarely observed. The author's field experience indicates that nonlayered coarse-grained gabbroic rocks occur where there is sufficient rock exposure as parts of thick tabular lava flows. Thomson (1950) and Lovell (1971) have shown on their geological maps many gabbroic bodies intruding the tholeiitic lavas. Recent remapping by the author shows that the gabbroic bodies are largely flows similar to those described by Pearce and Birkett (1974), and Jensen (1978a, 1978b). If some of the gabbroic bodies are sills, they have the same mineralogy as the thick lava flows. The following descriptions can be applied to all tholeiitic rocks in the Kirkland Lake area including the mafic parts of layered mafic-ultramafic sills (MacRae 1969; Naldrett and Mason 1968) in the komatiitic successions.

Magnesium-rich Tholeiitic Basalts

Magnesium-rich tholeiitic basalts are dark green to grey, or light green to grey, on weathered and fresh surfaces. These rocks occur as tabular flows, pillowed lavas, and pillow-breccias. The tabular flows are 1 to 100 m thick. The thicker flows can be traced along strike for several kilometres in areas where bedrock is well exposed. In places, the tabular flows can grade into pillowed lavas either along strike, or vertically.

The pillows are 30 to 150 cm in diameter and have dark green, deeply weathered selvages 1 to 3 cm thick. Pillowed flows form map units 10 to 600 m thick, which can be traced up to several kilometres where well exposed. The pillows may be closely packed, or they may be separated from one another by thick zones of hyaloclastite (Figure 15; Wells *et al.* 1979). In places, the pillows may be extensively fractured so that they can be described as pillow breccias. The fragments of pillows occur in a hyaloclastic matrix composed of fine-grained chloritized glass globules and shards. The pillow breccias are similar to those described by Henderson (1953), and Furnes (1972).

Hyaloclastite (0.1 to 5 m thick) also occurs at the top of many of the tabular flows (Jensen 1978a), and appears to have developed as a result of tachylite chipping off molten lava in contact with seawater (*see* Carlisle 1963; Furnes 1974; Dimroth *et al.* 1978; Sigvaldason 1968). An extensive description of the hyaloclastite from within the study area has been made by Dimroth and Lichtblau (1979).

Magnesium-rich tholeiitic basalt consists of 30 to 40% augite and 50 to 60% plagioclase (An_{55} to An_{70}). Magnetite, ilmenite, and sulphide grains form the remaining 1 to 3% of the rock. Some thick flows contain sparse enstatite phenocrysts, 0.5 to 1 cm in size altered to bastite, antigorite, and talc and, in other flows light, yellowish green to white phenocrysts of plagioclase, 0.5 to 3 cm across are present (Pearce and Birkett 1974; Jensen 1978a). The feldspar phenocrysts can form 0.5 to 5% of the rock, and are seen throughout the flow, including the fine-grained top and selvage; this suggests that the phenocrysts were formed at depth and then transported to surface. Flows with feldspar phenocrysts similar to those above have been described in other greenstone terrains by Green (1975), and in ocean-floor basalts with similar chemical compositions by Langmuir *et al.* (1977) and Rhodes *et al.* (1979).

Grains of plagioclase, clinopyroxene, and magnetite in the flow-tops and near the margins of pillows are small, and have spherulitic or dendritic habits. In some pillow lavas, the spherulitic feldspars form

TABLE 2: CHEMICAL ANALYSES OF THOLEIITIC BASALTS.

Major Components in %	Magnesium-rich Tholeiitic Basalts				Iron-rich Tholeiitic Basalt			
Sample Number	1	2	3	4	1	2	3	4
Field Number	72D536	73J1089	72J507	72J505	72J576	72D11	74J1237	71J581
SiO_2	46.60	47.70	48.10	47.70	47.70	54.10	47.20	50.60
Al_2O_3	14.50	15.70	15.80	17.30	13.00	11.30	14.50	12.20
Fe_2O_3	1.74	1.62	12.1*	10.4*	4.76	4.03	18.30	4.65
FeO	9.60	9.25	-	-	13.90	10.60	-	10.70
MgO	9.29	6.39	9.97	9.76	5.85	3.89	4.61	3.81
CaO	10.50	11.70	9.86	8.97	6.65	6.42	9.70	8.72
Na_2O	1.70	2.18	1.56	1.43	2.40	2.56	1.81	2.59
K_2O	0.17	0.48	0.10	0.63	0.39	0.15	0.17	0.38
TiO_2	0.83	1.11	1.02	0.73	2.36	2.35	2.36	2.46
P_2O_5	0.05	0.09	-	-	0.16	0.41	-	0.10
S	0.01	0.01	-	-	0.07	0.10	-	0.17
MnO	0.20	0.20	-	-	0.27	0.25	0.25	0.23
CO_2	0.10	1.00	-	-	0.19	0.20	-	0.14
H_2O^+	3.20	1.72	-	-	2.97	3.45	0.30	2.26
H_2O^-	0.08	0.21	-	-	0.06	0.09	-	0.32
TOTAL:	98.40	99.40	98.51	96.92	100.70	99.90	99.20	99.30

Notes:
*Total Fe reported as Fe_2O_3

Molecular Norms* Sample Number	1	2	3	4	1	2	3	4
Apatite	0.110	0.197	-	-	0.356	0.930	-	0.225
Ilmenite	1.218	1.618	1.459	1.054	3.497	3.547	3.513	3.686
Orthoclase	1.059	2.971	0.067	3.861	2.454	0.961	1.074	2.290
Albite	16.078	14.378	14.378	13.303	22.922	24.900	17.365	25.011
Anorthite	33.120	33.119	36.779	40.342	25.056	20.481	33.069	22.164
Magnetite	1.916	1.773	2.705	2.416	4.293	4.361	4.312	4.453
Enstatite	14.387	7.328	24.301	25.150	15.439	9.412	10.171	6.711
Ferrosilite	7.171	4.957	8.985	8.230	16.243	0.430	11.623	7.096
Quartz	-	-	-	-	2.336	16.705	4.180	8.790
Diopside	11.437	12.897	7.856	3.483	3.482	4.445	6.857	9.205
Forsterite	5.184	3.515	0.020	0.771	-	-	-	-
Fayalite	2.584	2.378	0.007	0.252	-	-	-	-
Hedenbergite	5.700	8.725	2.095	1.140	3.664	4.454	7.835	9.773

Notes:

1. Greenish-grey, aphanitic, magnesium-rich tholeiitic basalt, Frecheville Township (Analysis 62, Appendices II and III)
2. Dark green, aphanitic, magnesium-rich tholeiitic basalt, Lamplugh Township (Analysis 215, Appendices II and III).

3. Grey, 4 mm grained, magnesium-rich tholeiitic basalt, Thackeray Township (Analysis 167, Appendices IV and V).
4. Grey, massive, 2 to 3 mm grained, magnesium-rich tholeiitic basalt with plagioclase phenocrysts, Thackeray Township.

*Barth-Niggli Katanorm (Chayes and Metais 1963)

1.Massive, dark green, 1 to 2 mm grained, iron-rich tholeiite, Frecheville Township (Analysis 135, Appendices II and III).
2. Massive, dark green, 1 mm grained, iron-rich tholeiite Marriott Township (Analysis I, Appendices II and III).
3. Massive, black, 1 mm grained, iron-rich tholeiitic basalt, Cook Township.
4. Aphanitic, pillowed, iron-rich tholeiitic basalt, Elliott Township (Analysis 177, Appendices IV and V).

TABLE 3: CHEMICAL COMPOSITION OF THOLEIITIC DACITE AND RHYOLITE.

Major Components in Weight % Sample Number	1	2	3	4
Field Number	74F374	74F464	71J600	72J604
SiO_2	65.30	66.10	75.60	90.10
Al_2O_3	12.10	12.30	10.60	5.46
Fe_2O_3	9.20	9.10	0.91	0.79
FeO	-	-	2.79	-
MgO	1.15	0.77	0.69	0.01
CaO	3.06	4.38	1.30	0.78
Na_2O	5.20	4.39	3.85	0.72
K_2O	0.29	0.27	1.04	0.72
TiO_2	0.89	0.84	0.26	0.08
P_2O_5	-	-	0.03	-
S	-	-	0.01	-
MnO	0.11	0.17	0.05	-
CO_2	-	-	0.58	-
H_2O^+	1.15	0.15	0.69	-
H_2O^-	-	-	0.28	-
TOTAL	98.40	98.50	98.90	99.30

Molecular Norms* Sample Number	1	2	3	4
Apatite	-	-	0.666	-
Ilmenite	1.304	1.228	0.380	-
Orthoclase	1.803	1.675	6.460	4.549
Albite	49.085	41.348	36.3019	7.480
Anorthite	9.277	13.706	6.568	7.208
Magnetite	2.627	2.567	0.999	-
Corundum	-	-	.947	1.078
Enstatite	2.483	1.372	2.000	-
Ferrosilite	5.287	4.447	3.539	-
Quartz	22.784	26.382	42.677	79.037
Diopside	1.710	1.716	-	-
Hedenbergite	3.640	5.560	-	-

Notes:
1. Dark grey, fine-grained tholeiitic dacite, Barnet Township.
2. Black, fine-grained tholeiitic dacite, Barnet Township.
3. Light reddish-grey, aphanitic tholeiitic rhyolite, Elliott Township.
4. Light grey, cherty tuff of tholeiitic rhyolite composition, Thackeray Township.

*Barth-Niggli Katanorm (Chayes and Metais 1963)

variolites 0.1 to 1.5 cm in size. Downward in the tabular flows and inward in the pillows, the plagioclase and pyroxene grains become acicular and form intrafasiculate textures. In the thick tabular flows, greenish-grey ophitic augite grains (2 to 5 mm across), enclose many of the smaller plagioclase laths to give the rocks a gabbroic texture. These observations are consistent with the experimental studies of Lofgren (1974), Lofgren *et al.* (1974), Nabelek *et al.* (1978), and Walker *et al.* (1978) that relate the crystal habit to cooling rates. The lava flows, after being extruded almost phenocryst-free at temperatures of 1200 ± 100°C, cooled at rates greater than 60°C per hour in their outer parts to form glassy selvages. As the intrafasciculate textures formed, the rate of cooling inward from the glassy margin was about 6°C per hour. Thus, the interiors of most pillows appear to have cooled evenly at a rate of about 6°C per hour whereas the thicker tabular flows with gabbroic textures probably had interiors which cooled at a rate of less than 1°C per hour as their outer parts cooled at a faster rate.

Iron-rich Tholeiitic Basalt and Tholeiitic Andesite

Iron-rich tholeiitic basalt and tholeiitic andesite are black to dark green, magnetic rocks which weather to rusty brown. These rocks occur as tabular flows and pillowed lavas and, in a few places, as pillow-breccias. In most respects the rocks are similar to magnesium-rich tholeiitic basalts, except for their dark colours. The iron-rich tholeiitic basalts have dark green to

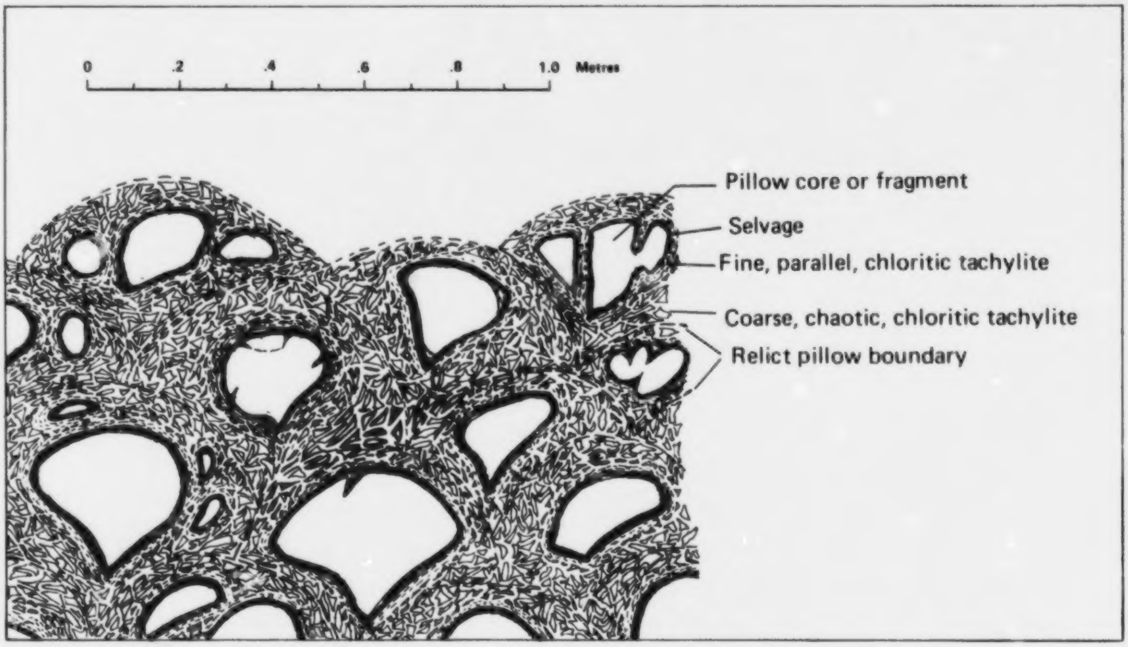

0
.2
.4
.6
.8
1.0 Metres
Pillow core or fragment
Selvage
Fine, parallel, chloritic tachylite
Coarse, chaotic, chloritic tachylite
Relict pillow boundary

Figure 15. Diagram of pillow lavas separated by hyaloclastite.

black pyroxene and abundant titaniferous magnetite which give them their colour and magnetic properties.

Hyaloclastite (0.1 to 10 m thick), occurs at the top of many tabular flows (Figure 16). It is similar to that found above flows of magnesium-rich tholeiitic basalt.

Iron-rich tholeiitic basalt consists of 40 to 50% augite, 30 to 40% plagioclase (An_{40} to An_{60}), 5 to 10% titaniferous magnetite, and 2 to 8% quartz. In places, hornblende, pseudomorphic after augite, occurs in the iron-rich tholeiitic basalt (Pearce and Birkett 1974; Jensen 1978a). Tholeiitic andesites consist of 50 to 60% plagioclase (An_{25} to An_{45}), 30 to 35% hornblende, 5 to 10% titaniferous magnetite, 10 to 20% quartz, and 1 to 3% apatite and sulphide minerals.

In the iron-rich tholeiitic basalts, the crystal habits of pyroxene, plagioclase, and titaniferous magnetite are similar to the crystal habits in the magnesium-rich tholeiitic basalts. This suggests the lavas of both groups of rock had similar cooling rates.

Tholeiitic Dacite and Rhyolite

Tholeiitic dacite and rhyolite cannot be distinguished from each other except by chemical analyses. These rocks form units 3 to 100 m thick which have strike lengths of 1.5 to 10 km. These rocks are cherty, siliceous rocks, mainly devoid of quartz and feldspar phenocrysts. They are generally reddish grey to reddish brown but can range from black to light yellow, and have a mottled appearance (Jensen 1978a).

In the field, the association with iron-rich tholeiitic basalt distinguishes them from calc-alkalic dacites and rhyolites. The general lack of quartz phenocrysts in the tholeiitic dacites and rhyolites suggests that these rocks, like the iron-rich tholeiitic basalts of D.H. Green and Ringwood (1967a, 1967b), were differentiated from tholeiitic magmas at depths of less than 15 km.

The mineralogy and textures found in the tholeiitic dacites and rhyolites have been described by Jensen (1978a). Many of these rocks are composed of orbs (Lofgren 1971a) 0.1 to 0.5 mm in diameter that are made up of quartz and spherulitic albite grains 0.002 to 0.005 mm in size. The orbs are rimmed by pyrite, magnetite, chlorite, and in places, pumpellyite. The orbs result from the devitrification of rhyolitic glass (Lofgren 1971a). In other places, the dacites and rhyolites have a granular texture of interlocking quartz, calcite, and albite grains, 0.01 to 0.05 mm in maximum dimension, associated with finely disseminated chlorite, pyrite, and magnetite grains, 0.002 to 0.005 mm in maximum dimension.

CALC–ALKALIC VOLCANIC ROCKS AND RELATED INTRUSIVE ROCKS

RECENT STUDIES

There is little doubt that calc-alkalic volcanic rocks represent a suite of rocks distinctive from suites of tholeiitic and alkalic volcanic rocks (D.H. Green and Ringwood 1967a, 1967b; Masuda and Aoki 1979). What is still uncertain is whether or not calc-alkalic rocks are solely products of partial melting of the upper mantle under hydrous conditions as proposed by O'Hara (1965), Yoder (1969), Kushiro *et al.* (1972), Kushiro and Yoder (1973), Kushiro (1974). Nicholls and Ringwood (1972) have argued against such a process, and favour a 2-stage model whereby basaltic crust is subducted, dehydrated, and transformed into quartz eclogite which is then partly melted to produce calc-alkalic magmas (T.H. Green and Ringwood 1968). Refinements on this model have

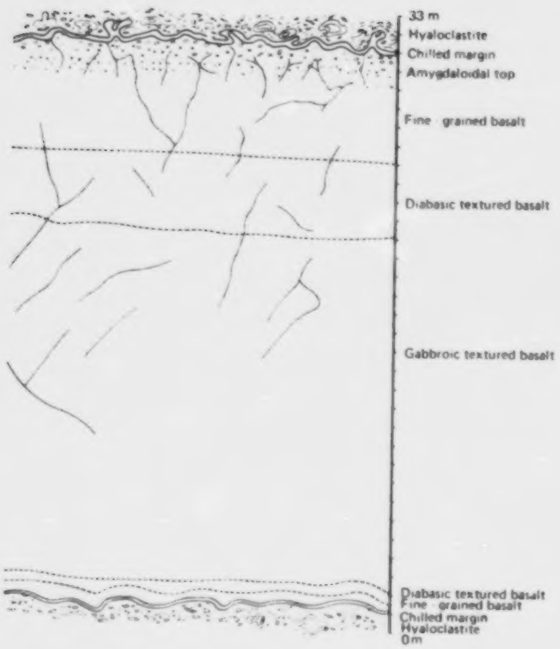


Figure 16. Diagram of a massive iron-rich tholeiitic basalt flow. Multiply m by 3.21 to obtain feet (after Jensen 1976b).

been made by Nicholls and Ringwood (1973), Ringwood (1974), Nicholls (1975) whereby partial melts of subducted ocean crust are fused with the overlying mantle to form wet pyroxenitic diapirs. The diapirs upon rising, partly melt at various depths to produce a range of calc-alkalic magmas. A third method for producing calc-alkalic magmas is the partial melting of crustal rocks at the base of island arc volcanic piles and continental margins (Jakes and White 1971). However, most of the research on calc-alkalic volcanic rocks suggests that they have a magma source which was generated in the upper mantle but which was not in equilibrium with ultramafic material (Stern 1974).

Gill (1974) emphasized that the REE patterns of Fijian calc-alkalic rocks were incompatible with eclogite equilibria. This observation has been confirmed for most other modern island-arc calc-alkalic rocks (for example, Whitford *et al.* 1979; Masuda and Aoki 1979), whereas the opposite appears to be true for Archean calc-alkalic volcanic rocks (Nesbitt and Sun 1976). Other authors (Arth and Hansen 1972, 1975; Jahn *et al.* 1974; Condie and Baragar 1974; Condie and Harrison 1976; Condie 1975; Glikson 1976a; Hawkesworth and O'Nions 1977; and O'Nions and Pankhurst 1978; and Ewing 1979); have concluded that partial melting of eclogite or amphibolite similar in composition to tholeiitic basalt, was responsible for the depletion of the HREE found in the Archean calc-alkalic volcanic rocks of their studies. This suggests that unlike younger calc-alkalic rocks, Archean calc-alkalic rocks were not generated in the upper mantle. This would also indicate that Archean calc-alkalic rocks were not generated above Benioff seismic zones.

Archean calc-alkalic rocks have Ni and Cr contents which are too high for them to have the same origin as Cenozoic island arc volcanic rocks (Jolly 1975; Hawkesworth and O'Nions 1977; Nesbitt and Sun 1976; Glikson 1976a; Taylor 1979).

Condie (1976a, 1976b), Smith (1978), and Ewing (1979), have pointed out that Archean calc-alkalic dacites and rhyolites compared to basalts and andesites tend to have lower REE abundances and steeper REE patterns which show strong HREE depletion. Ewing (1979) concluded that the Archean basalt-andesite patterns were similar to those of calc-alkalic basalts and andesites found in continental environments and that these patterns were controlled in varying amounts by amphibole, garnet, and pyroxene fractionation. As for the dacites and rhyolites, he concluded that they had no modern equivalent, and that their patterns were probably derived by the partial melting of eclogite having a composition similar to andesite. A similar conclusion was made by Smith (1978). Furthermore, Ewing (1979) and Smith (1978) both confirmed the findings of Arth and Hanson (1975) that the calc-alkalic dacites and rhyolites had REE patterns that corresponded to those of Archean tonalitic intrusions.

NOMENCLATURE AND CLASSIFICATION

The term 'calc-alkalic' was introduced by Peacock (1931). Historical reviews of the studies leading to the introduction of the term, and of subsequent studies concerning the term's present meaning, have been given by Turner and Verhoogen (1960), Barth (1962), Wilkinson (1967), and Irvine and Baragar (1971). The most distinctive features of calc-alkalic basalts were their higher Al_2O_3 contents and their lower FeO/MgO ratios relative to tholeiites with similar silica contents, and their close field association with more evolved calc-alkali lavas (Wilkinson 1967). The problems of distinguishing calc-alkalic basalt from tholeiitic basalt and of dividing the calc-alkalic rocks into basalt, andesite, dacite, and rhyolite have been discussed by Irvine and Baragar (1971). These authors suggested that the AFM diagram and the Al_2O_3 versus normative plagioclase composition plot be used to distinguish tholeiitic from calc-alkalic rocks. They further recommended that the normative colour index versus normative plagioclase composition plot be used to differentiate between basalt, andesite, dacite, and rhyolite rocks. Church (1975) and Jensen (1976a) discussed the problems associated with the methods proposed. To avoid the problems of alkali, silica, and calcium alteration in classifying calc-alkalic rocks, Jensen (1976a) suggested using a ternary diagram involving the cation percentages of Al_2O, Fe_2O_3 + Fe_2O + TiO_2, and MgO, as a means of distinguishing between the tholeiitic and calc-alkalic rocks, and further dividing the calc-alkalic group into basalt, andesite, dacite, and rhyolite (*see* Figure 7). This cation diagram, unlike the method given by Irvine and Baragar (1971), has the advantage of distinguishing between tholeiitic and komatiitic rocks (Figure 17).

FIELD AND PETROGRAPHIC CHARACTERISTICS OF CALC-ALKALIC VOLCANIC ROCKS

Calc-alkalic volcanic rocks in the Kirkland Lake area have a wider range of physical characteristics than do the komatiitic and tholeiitic volcanic rocks. These rocks are considered to be products of volcanism that ranged from quiet lava effusions to violent explosive eruptions of fragmented lava. In addition, these lavas were probably erupted from cone volcanoes with steep slopes and variable surfaces, which would have affected the morphology of the lava flows and fragmental rocks. Lavas, particularly the fragmental types, deposited on the upper parts of volcanic edifices would also be subjected to slumping and redeposition away from their eruptive centres. Therefore, the physical characteristics of the calc-alkalic lavas are important to the interpretation of the history of the area.

At this point, only the general field features of calc-alkalic rocks and their petrography will be discussed. Field features particular to parts of the Kirkland Lake area will be included in the discussions concerning the stratigraphy.

Calc-alkalic Basalts and Andesites

Descriptions of these rocks have been made for the Kirkland Lake area by Jensen (1975a, 1975b, 1978a), and for the Noranda area, Quebec, by Dimroth *et al.* (1973), Dimroth *et al.* (1978), Dimroth and Rocheleau 1979). Calc-alkalic basalt and andesite cannot be distinguished readily from one another in the field except by slight changes in colour and texture. The 2 types are light green to grey, interlayered, and in many places grade imperceptibly into one another, which makes accurate field delineation impossible. Consequently, these rocks have been mapped according to physical forms such as types of flows and fragments (Jensen 1978a).

The flows are tabular and are from 1 to 35 m thick, and have strike lengths of 0.5 to 3 km. The flows can occur as a continuous series of flows, or they can be interlayered with pillow lavas. Dark green chloritic chill margins, 2 to 3 cm thick, mark the boundaries of the flows.

Pillowed lavas, 0.1 to 1 m across, form map units 10 to 350 m thick, having strike lengths of 0.5 to 10 km. Many of these map units are lenticular. Pillows may be closely packed with small amounts of subangular volcanic fragments, quartz, calcite, and chlorite in the spaces between the pillows; or they may be loosely packed and separated by a light to dark green matrix composed of subangular volcanic fragments (1 to 10 mm in size), quartz and feldspar grains (0.1 to 1 mm), and chloritized glass fragments (1 to 10 mm) set in a mesostasis of chlorite, albite, quartz, and epidote. The source of matrix material is thought to be the pillows themselves and earlier formed tuffaceous material which may have slumped down slope with the pillows.

The mineral composition of the calc-alkalic basalts and andesites are similar. Fine-grained varieties consist of 10 to 25% tabular phenocrysts of zoned plagioclase (An_{35} to An_{69}), 0.5 to 2.0 mm long, which occur in clusters 2 to 4 mm in size, to produce a glomeroporphyritic texture. The phenocrysts occur in a felty groundmass of acicular to tabular grains of plagioclase, 0.01 to 0.05 mm long, with disseminated equant to subequant augite grains 0.005 to 0.02 mm in size, and minor acicular to equant grains of magnetite and pyrite. Quartz, chlorite, and epidote occur interstitially to the plagioclase and pyroxene in the groundmass.

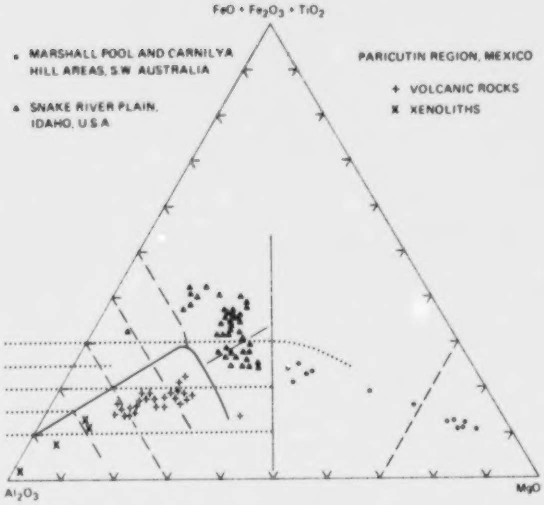


Figure 17. Jensen Cation Diagram comparing the patterns of variation of komatiitic, tholeiitic, and calc-alkalic rock suites. Solid lines serve to separate komatiitic, tholeiitic, and calc-alkalic rock suites, dashed lines separate rock types, dotted lines serve as colour boundaries (*see* Figure 7). Oxides are calculated in cation percent (after Jensen 1976a).

In places, the basalt and andesite rocks have up to 20% hypersthene phenocrysts 0.5 to 2.0 mm long, set in the felt plagioclase groundmass. The hypersthene and plagioclase appear to have formed prior to the extrusion of the magmas.

Where the fine-grained rocks grade into coarser grained rocks toward the central parts of tabular flows, equant grains of ophitic augite, 1 to 2 mm in size, enclose zoned tabular plagioclase grains 0.5 to 3.0 mm long, and the porphyritic texture of the rock becomes less evident. Augite forms 15 to 35% of the rock, and plagioclase, forms 60 to 70%. Interstitial quartz and disseminated magnetite and pyrite grains 0.1 to 1.0 mm form the remainder of the rock.

Calc-alkalic Dacites and Rhyolites

These rocks have been described in the Kirkland Lake area by Ridler (1969) and Jensen (1975a, 1975b), and in the Noranda area by DeRosen Spence (1976), Dimroth *et al.* (1975), Dimroth and Demarcke (1978), Dimroth *et al.* (1979).

The calc-alkalic dacite rocks are difficult to distinguish from rhyolite without chemical analyses. The dacites tend to have fewer and smaller quartz phenocrysts than rhyolite. The 2 rock types occur together in subvolcanic intrusions and in fragmental extrusive map units. The rock types are generally light grey, green, pink, or white and where massive, are hard, siliceous rocks. Where brecciated or foliated, these rocks contain sericite and are less hard. The fragmental extrusions are flow-breccias or pyroclastic rocks that range from tuff-breccias with coarse angular fragments, to fine-grained layered tuffs composed of lithic, crystal, and ash fragments cemented by chert.

The dacites and rhyolites are porphyritic rocks that contain plagioclase and quartz phenocrysts 0.5 to 5 mm in size, set in a matrix of quartz, albite, sericite, chlorite, and pyrite grains 0.001 to 0.02 mm in size. The plagioclase phenocrysts (An_{15} to An_{30}) form 5 to 30% of the rocks, and are zoned tabular grains that vary from vitreous to milky white depending on alteration. The quartz phenocrysts form 2 to 10% of the rocks and are clear, euhedral quartz grains, many have a bipyramidal shape, some have rounded corners and deep embayments filled with matrix material.

The colour and size of the plagioclase and quartz phenocrysts tend to be distinctive in each of the 3 groups of calc-alkalic rock; the Skead, Blake River, and Hunter Mine Groups (*see* Figure 5). This feature makes the plagioclase and quartz phenocrysts important for recognizing the 3 groups of rock.

T.H. Green and Ringwood (1968) concluded that the formation of quartz phenocrysts in dacitic magmas, derived from the partial melting of quartz eclogite, required pressures equivalent to depths of about 100 km. At shallower depths, plagioclase would replace quartz as the liquidus phase and resorption of the quartz phenocrysts would take place as the magma was being transported to surface.

Intrusive Rocks

Intrusive rocks composed of gabbro, quartz gabbro, hornblende gabbro, diorite, and quartz diorite cut the calc-alkalic volcanic rocks of the Blake River Group. These rocks form elongate to circular bodies less than 3 km across. Only a few of the bodies have 2 or more phases of gabbro and diorite, and where they do so, the change is gradual and difficult to recognize. Most of these intrusive bodies have been individually described by Jensen (1972, 1975a, 1975b, 1978a).

All of the intrusive rocks are medium grained (2 to 3 mm in size), except at their contacts with the volcanic rocks, where they become finer grained than is normal. Gabbro, the most abundant of the intrusive rock types, has an ophitic texture, and consists of 40 to 50% augite, 30 to 40% saussuritized plagioclase, 2 to 3% magnetite, and as much as 2% quartz. Quartz gabbro and hornblende gabbro are similar to the gabbro, except that they have 5 to 15% quartz, and the chief mafic mineral in hornblende gabbro is hornblende rather than augite.

Diorites are light grey to green, the chief mafic mineral being hornblende, but augite can be the dominant mafic mineral in some. The diorites consist of 60 to 65% plagioclase (An_{35} to An_{50}), 25 to 35% subhedral hornblende or augite, 0 to 10% quartz, and 2 to 3% magnetite. Biotite and apatite occur as accessory minerals.

Quartz diorite is grey to pink granitic rock composed of 60 to 65% plagioclase (An_{30} to An_{40}), 15 to 25% subhedral hornblende, 10 to 15% quartz and 2 to 3% magnetite and biotite.

The range of chemical compositions of the intrusions is similar to those of the calc-alkalic volcanic rocks within the Blake River Group. There are also close spatial relationships between gabbro and calc-alkalic basalt and andesite, diorite and quartz diorite, and calc-alkalic dacite and rhyolite, to suggest that the intrusive rocks and the volcanic rocks were closely related to each other at the time of their formation.

MASSIVE TO GNEISSIC QUARTZ DIORITE, TONALITE, AND TRONDHJEMITE

The Lake Abitibi Batholith, the Round Lake Batholith (Figure 5), and possibly, the quartz diorite stocks that cut the Blake River Group are the only rocks in the Kirkland Lake area which belong to this group of granitic rocks. These batholiths and stocks of massive to gneissic quartz diorite, tonalite, and trondhjemite are considered by the authors to be similar to Na-rich granitic rocks found elsewhere in Archean terrains where they are sometimes referred to as calc-alkalic tonalites (Tarney and Windley 1977). In this section, these rocks will be referred to as Na-rich granites or granitic rocks.

Gneissic belts containing granulites and anorthosites are 85% 'calc-alkalic' quartz feldspathic gneisses of tonalitic composition cut by undeformed tonalites and granodiorites (Windley 1979), and appear to represent deep erosional levels of Archean crustal rock (O'Hara 1977). Recent studies suggest that the presence of granulites and anorthosites, in terrains with mainly Na-rich granitic rocks, could represent important clues as to the origin of Na-rich granites found within the study area, where only 2 to 3 km of erosion occurred (Jolly 1978).

RECENT STUDIES CONCERNING ARCHEAN Na-RICH GRANITIC ROCKS

Almost all of the studies on the isotopic and trace element contents of Archean Na-rich granitic rocks suggest that these rocks represent new additions of sialic material to the earth's crust (Moorbath 1976). Most of these data also suggest that Na-rich granites must represent the partial melting of either quartz eclogite or garnet-bearing amphibolite of mafic to intermediate composition at depths of 45 to 150 km (Hanson and Goldich 1972; Glikson 1972, 1976a, 1979b; Moorbath 1975a; Arth and Hanson 1975; O'Nions and Pankhurst 1974, 1978; Arth and Barker 1976; and Hanson 1978). Spooner and Fairbairn (1970), D.C. Green and Baadsgaard (1971), Hawkesworth *et al.* (1975), Moorbath (1976), Barker and Arth (1976) have related the above process to the formation of calc-alkalic andesites, whereas O'Nions and Pankhurst (1974) and Moorbath (1975a) have suggested that andesitic rocks were the precursors of Na-rich granitic rocks.

The validity of this interpretation has been challenged by Shaw (1976), Bridgwater and Collerson (1976), O'Hara (1977), Taylor and Hallberg (1977), and Fyfe (1978a, 1978b). These authors have suggested that the Sr and Pb isotope ratios and Rb contents of the Na-rich granites have undergone mantle imprinting during the metasomatism of pre-existing sialic rock (crust). Collerson and Fryer (1978a, 1978b) have also suggested that the REE's have been affected by hydrothermal fluids during granulite facies metamorphism and deep crustal anatexis of sialic crust. These authors also proposed that K-rich granites represented anatectic melts generated under granulite or amphibolite facies conditions from previously formed 'plagiogranites'. These proposals by Collerson and Fryer (1978) have been challenged by Glikson (1979b).

Moorbath (1975a) pointed out that the ancient gneisses must have been derived from calc-alkalic igneous rocks. Since metamorphic processes themselves probably have little, if any effect on REE patterns, the ancient gneisses must have inherited their REE patterns from their precursory calc-alkalic rocks. Archean gneisses and tonalites have REE patterns which correspond more closely to calc-alkalic dacites and rhyolites than to calc-alkalic basalts and andesites (Ewing 1979). The calc-alkalic dacites and rhyolites required to form granitic rock by metamorphism are not abundant and tend to be restricted to the upper parts of calc-alkalic volcanic successions (Goodwin 1968, 1979). It is unlikely that calc-alkalic dacites and rhyolites could be metamorphosed without remelting the underlying calc-alkalic andesites and basalts. Thus, it is more likely that the Na-rich granites were emplaced at depth during felsic calc- alkalic volcanism.

Sources of 'intermediate' composition are required for rhyolitic and Na-rich granitic magmas (Hawkesworth and O'Nions 1974; Smith 1978; Ewing 1979). Calc-alkalic basalt and andesite could provide the required sources provided that these rocks were transformed into amphibolite or garnet bearing rocks (eclogite), and were partly melted. Such a process would require either that the calc-alkalic volcanic successions be very thick, or that the calc-alkalic basalts and andesites be effectively subducted to subcrustal depths. If deeply buried calc-alkalic basalts and andesites were converted to amphibolite and garnet-bearing amphibolite which then underwent partial melting, the resulting magma would have been of andesitic composition. Andesitic magmas occur in a thermal trough between more basic and more felsic magmas (D.H. Green and Ringwood 1967a, 1967b). Such magmas would tend to fractionate basic plagioclase with subordinant pyroxene and ore minerals to form anorthositic cumulates. The resulting lower melting acid magmas would have migrated upward by filter pressing (T.H. Green 1969). Such a process could explain the presence of granulites and anorthosites in gneissic, more deeply eroded Archean terrains cut by Na-rich granites. According to Tarney and Windley (1977), anorthosites tend to show LREE enrichments, strong HREE depletions, and strong positive Eu anomalies. These authors suggested that there was a genetic relationship between the anorthosites and associated calc-alkaline plutons, gneisses, and granulites. Tarney and Windley (1977) noted that because of the low abundance of anorthosite, the positive Eu anomalies in the anorthosites only partly compensated for the negative Eu anomalies in the higher granitic rocks. These authors concluded, therefore, that refractory cumulates with positive Eu anomalies (and rich in HREE) must have been carried down into the mantle.

Calc-alkalic andesites tend to be rare or absent in most greenstone belts (Barker and Peterman 1974; Glikson 1976a; Hubregtse 1976). Arth and Barker (1976) and Barker and Arth (1976) suggested that: andesitic magmas were probably formed at depth by the partial melting of quartz eclogite of basaltic composition; and that these magmas, at higher levels, must have undergone plagioclase and amphibole fractionation directly, to produce magmas of Na-rich granitic and calc-alkalic dacitic and rhyolitic composition. Cumulates of basic plagioclase, hornblende, and pyroxene would then result (Barker and Arth 1976), which could also explain the presence of anorthosite and mafic granulites in deeply eroded Archean gneissic terrains.

The process proposed by Barker and Arth (1976) would cause andesites, where they occurred, to have REE patterns similar to those of calc-alkalic dacites and rhyolites instead of the 2 distinct REE patterns described by Ewing (1979). To produce these 2 distinct patterns as well as the REE patterns in anorthosites observed by Tarney and Windley (1977) would require a part of the andesite magma to have recrystallized, forming a garnet-bearing cumulate. This would then have to sink back down into the mantle, after a part of the magma had been tapped to produce andesite lavas.

The similarity of Archean Na-rich granites to Cordilleran granitic rocks in their composition, and particularly in their occurrence as large composite batholiths, has led many scientists to propose that Archean granitic rocks were formed largely in tectonic environments similar to those represented by continental margin orogenic fold belts (for example Windley 1977b). Many have suggested that large batholiths such as the Sierra Nevada Batholith mainly represent reworked sialic crust and the anatexis of sedimentary rocks (for example, Bateman and Eaton 1967). However, Brown (1977) has argued that such granitic rocks have strontium isotopes which are inconsistent with crustal fusion, and, instead, he proposed that the magma for such rocks must come from the mantle ia manner similar to that of magma for calc-alkalic volcanic rocks above Benioff zones.

The isotopic and trace element data suggest that the magmas for Archean Na-rich granitic rocks were derived directly from the partial melting of garnet-bearing amphibolite, or eclogite similar in composition to basalt and andesite (Hawkesworth and O'Nions 1977; Moorbath 1975a). This would tend to preclude a tectonic environment, involving a Benioff zone as well as a process involving the reworking of sialic crust. However, the data do not preclude a tectonic environment which involved the thickening of basic to intermediate rocks above the mantle, either by volcanic deposition, or, as suggested by Tarney and Windley (1977), by thrusting slabs of such material on top of one another at the edge of a continental margin.

NOMENCLATURE AND CLASSIFICATION

The granitoid rocks of the map area are classified in this report according to the scheme described by Streckeisen (1976). In the map area, the granitic rocks are

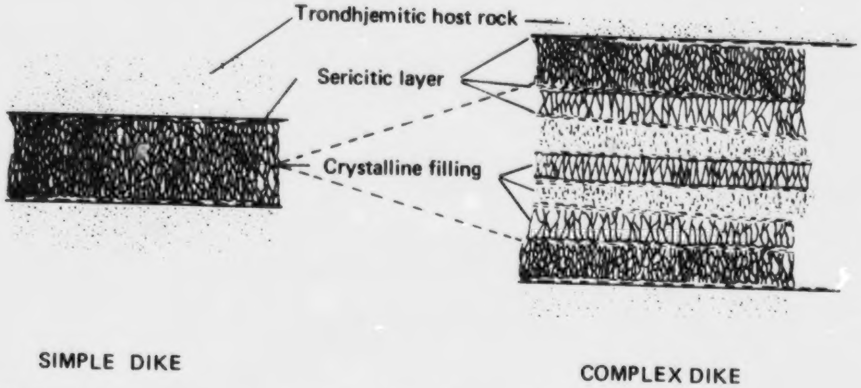


Figure 18. Illustration of the pegmatite dikes found in the trondhjemite rocks.

distinguished as Na-rich granitic and K-rich granitic rocks. The Na-rich granitic rocks are mainly quartz diorite, tonalite, and trondhjemite which comprise the Lake Abitibi Batholith and the Round Lake Batholith, and a few previously described granitic stocks in the Blake River Group. The K-rich granitic rocks include mainly syenite, monzonite, quartz monzonite, granodiorite, and quartz monzodiorite in which K-feldspar generally forms more than 25% of the total feldspar. The K-rich granitic rocks form stocks and dikes of equigranular to porphyritic rock and are possibly related to the alkalic volcanic rocks. Consequently the K-rich granitic rocks are described elsewhere in the report with the alkalic volcanic rocks and only the Na-rich granitic rocks are included in this part of the report.

FIELD AND PETROGRAPHIC FEATURES

In general, Na-rich granitic rocks are represented by sparse, low-lying, rounded, well jointed outcrops. These rocks are well exposed only along the shorelines of lakes. A prime example of such rocks is the granitic rocks of the Lake Abitibi Batholith.

In the Lake Abitibi Batholith, the rocks are mainly massive to gneissic trondhjemites. Lesser amounts of gneissic tonalite and quartz diorite occur toward the margins of the batholith. Many pegmatitic dikes of trondhjemitic composition cut the granitic rocks.

The massive trondhjemites are leucocratic, hypidiomorphic granular rocks composed of 20 to 30% anhedral quartz grains, 0.5 to 2 mm across, and 60 to 70% subhedral zoned fresh plagioclase (An $_{10}$to An $_{30}$) grains, 1 to 4 mm long. Biotite, muscovite, microcline, magnetite, and, in a few places, hornblende, form the remaining 5 to 10% of the rocks.

In strongly gneissic trondhjemites, the feldspar grains form the cores of augen, 2 to 8 mm long, and are surrounded, particularly at the ends of grains, by recrystallized finger grained interlocking quartz, albite, and perthite, to form oval volumes enclosed by platy biotite grains 0.5 to 2.0 mm across. The plagioclase grains have oscillatory and gradation zoning, and are altered to saussurite and kaolinite. The biotite is partly altered to chlorite and iron oxide.

The tonalites are mesocratic to leucocratic gneissic rocks composed of: 50 to 60% zoned plagioclase (An_{60} to An_{30}) grains 2 to 6 mm long, 10 to 20% quartz grains 1 to 4 mm across, and 10 to 15% hornblende 1 to 4 mm long. Magnetite, biotite, and perthite form the remaining 5 to 10% of the rocks. In places, the chief mafic mineral is biotite instead of hornblende. The augen gneissic texture in the tonalites is similar to that of the gneissic trondhjemites, except that the quartz-feldspar augens are more strongly separated by the mafic minerals, and hornblende replaces biotite as chief mafic mineral. In most places, the hornblende and biotite are partly altered to chlorite and the anorthite-rich cores of the plagioclase grains are saussuritized.

The quartz diorites are mesocratic to melanocratic augen gneisses. Light coloured areas, 2 to 8 mm across, composed of interlocking quartz and plagioclase (An_{70} to An_{30}) grains 1 to 4 mm in size, form the augen which are enclosed by dark green prismatic hornblende grains, 1 to 3 mm long, and platy biotite grains 0.5 to 2 mm across. Biotite and hornblende as well as minor magnetite make up about 30 to 50% of the rock composition, and are generally mylonitized and altered to chlorite. The plagioclase grains have strongly developed oscillatory zoning and tend to form the largest grains in the rock. The plagioclase grains are saussuritized particularly at their more anorthite-rich cores. The outer edges of the quartz-feldspar areas which are mylonitized, are recrystallized to form finer grained quartz and albite.

Pegmatite dikes of trondhjemitic composition, 5 cm to 1 m wide cut all the rocks in the batholith, and in many places cut one another. Most have straight sharp contacts with their host rocks and appear to represent fracture fillings. Many dikes have thin micaceous layers that alternate with thicker layers of interlocking quartz and feldspar (Figure 18). The layers are oriented parallel to the strike of the dike while the quartz and feldspar grains are oriented perpendicular to the strike of the dike.

ALKALIC VOLCANIC ROCKS AND K–RICH INTRUSIVE ROCKS

ALKALIC VOLCANIC ROCKS IN THE KIRKLAND LAKE AREA

Alkalic volcanic rocks near Kirkland Lake, have been reported and described by Burrows and Hopkins (1925), Todd (1928), Cooke (1966), and Cooke and Moorhouse (1969). These rocks occur with Timiskaming sedimentary rocks which unconformably overlie Keewatin-type (komatiitic, tholeiitic, and calc-alkalic) volcanic rocks (Thomson 1950). Cooke (1966) described the alkalic volcanic rocks as being mainly trachytes, mafic trachytes and leucitites. In addition, Prest (1957) and Satterly (1949a, 1952b, 1953) reported trachytes and Timiskaming-like sedimentary rocks along the Destor - Porcupine Fault Zone north of Kirkland Lake. In both areas, numerous intrusions of syenite, monzonite, and quartz monzonite cut all the rocks, and suggest that the trachytic rocks are closely associated with these intrusive rocks. As well, in both areas, the alkalic volcanic rocks are located along zones of major faulting to which they may also be related.

Cooke (1966) concluded, on the basis of intrusive relationships, that most of the syenitic stocks could be much younger than the trachytic volcanic rocks. The radiometric ages obtained on the syenitic stocks are still inconclusive (Ridler 1977). Matachewan diabase dikes, dated at 2690 Ma by Gates and Hurley (1973), crosscut the Otto Syenite Stock dated at 2160 ± 80 Ma by Bell and Blenkinsop (1975).

Whether or not the syenites and trachytes were related to each other, would greatly affect the geological interpretation of the area. If the 2 rock types are considered to be associated, information gained about the magma sources and the tectonic environment of 1 type could be extended and applied to the other. On the other hand, if the 2 are unrelated, as suggested by Cooke (1966), then the 2 groups of rock could have had distinct magma sources and represent 2 separate tectonic environments.

The study by Cooke (1966) did not reveal the presence of K-rich or Na-rich alkali olivine basalt, ankaramite, or alkali basalt. Nevertheless, Ridler (1969) concluded that many of the volcanic rocks south of Kirkland Lake were alkali olivine basalts and ankaramites, and were, therefore, part of the Timiskaming Group. However, Jensen (1976b) concluded that Ridler's alkalic volcanic rocks were actually komatiites enriched in alkalies by metasomatic activity associated with the numerous syenitic stocks and dikes which cut these rocks. Ridler's alkalic rocks have relict spinifex textures and the polyhedral jointing of komatiites, low-TiO_2 contents atypical of alkali basalts, and become subalkalic further away from the syenitic intrusions. Extensive aureoles of potassic fenitisation are common around syenitic intrusions (Baldock 1971).

Goodwin (1979) concurred with Ridler's (1969) correlation of the Timiskaming Group with the mafic volcanic rocks south of Kirkland Lake. Goodwin (1979) also located similar alkali-rich mafic rocks in other places where he concluded that they were probably caused by local chemical alteration.

Alkali olivine basalt, alkali basalt, and nepheline related to the deposition of the trachytic volcanic rocks may be present along the southern part of the Destor - Porcupine Fault Zone (*see* Figures 5 and 6). Outcrop is sparse along this part of the fault zone. The main information available for this area comes from about 400 drill logs, and approximately 3000 short, but well located, drill core specimens collected in the 1940s by N. Hogg for the Ontario Government. A few chemical analyses and thin sections of these rocks suggest that trachytic rocks are abundant, and that a few high-TiO_2 alkali basalts and nephelinites could be present.

STUDIES CONCERNING ARCHEAN ALKALIC ROCKS

Alkalic Volcanic Rocks

The paucity of alkalic olivine basalt in Archean greenstone belts has been pointed out by Hallberg and Williams (1972), and Hawkesworth and O'Nions (1977). However, apart from having been found in the Kirkland Lake area (Cooke 1966), trachytic rocks have also been reported in the Barberton Greenstone Belt by Anhaeusser (1971b), and rocks of shoshonitic affinities have been reported in the North Spirit Lake Greenstone Belt of Northwestern Ontario by Smith and Longstaff (1974) as well as in the Knee Lake - Oxford Lake Greenstone Belt in Manitoba, by Hubregtse (1976). Cooke (1966) and Smith and Longstaffe (1974) favoured an island arc environment for the development of these rocks, Hubregtse (1976), however, concluded that an island arc environment was not required. Smith and Longstaff (1974) further concluded that their shoshonitic rocks were associated with the nearby plutons of adamellite, granodiorite, tonalite, and monzonite.

In general, Cenozoic alkalic volcanism that includes trachytic lavas appears to occur mainly on oceanic islands, along zones of continental crust rifting and extension, behind island arcs, and on the continental sides of orogenic belts. In many of these places, the detailed tectonic setting of the alkalic volcanism is still uncertain (Schwarzer and Rogers 1974; Cundari 1979; Gest and McBirney 1979). Rifting in both continental and oceanic crusts seems to be closely associated with alkalic volcanism (Best and Brimhall 1974; Villari 1974; Fitton and Hughes 1977).

The rocks studied by Cooke (1966) do not appear to represent ocean island volcanism, nor do they appear to represent volcanism that occurs along many continental rift zones. The leucitic trachytes, mafic trachytes, and trachytes of Cooke (1966) are trachybasalts, tristanites, and high-K calc-alkalic andesites and dacites, respectively, according to the classification of Irvine and Baragar (1971). There is a lack of alkalic olivine basalt and alkali basalt among the rocks and the rocks, with about 50% SiO_2 have about 1% TiO_2.

On oceanic islands, alkali olivine basalt and alkali basalt predominate over trachytic rocks. Where rocks with about 50% SiO_2 do occur, they have more than 2% TiO_2 (Engel *et al.* 1965; Baker 1974; Hawkins and Natland 1975; Sullivan and Keen 1977). In most zones of rifting and extension of sialic crust, even though trachytes may predominate, the rocks have high-TiO_2 contents as well (Nash *et al.* 1969; Kesson 1973; Noble and Parker 1974; Best and Brimhall 1974; Barberi *et al.* 1974; Ferguson and Cundari 1975; Baker *et al.* 1977; Kiselev *et al.* 1978).

The rocks of Cooke (1966) resemble many of the rocks described by Joplin (1968) as being shoshonitic lavas. High-K_2O mafic trachytes (shoshonites) with low-TiO_2 (<1%) tend to occur in island-arc environments (Jakes and White 1971; Jolly 1971; Mackenzie and Chappell 1972; Brown *et al.* 1977). These high-

K_2O alkaline lavas have been considered by many to be generated in areas where there is a Benioff seismic zone 250 to 400 km below surface (for example by Hatherton and Dickinson 1968; Whitford and Nicholls 1976). T.H. Green (1980) has further suggested that island-arc, high-K_2O alkaline suites are related to zones of subduction because of their low-TiO_2 content, and can be distinguished from other alkaline suites. This would tend to support the proposal by Cooke (1966) that the trachytic rocks of Kirkland Lake were formed in an island-arc environment.

Many high-K alkalic basalts of island arc environments have high-TiO_2 contents and are similar to alkali basalts of other environments (Schwarzer and Rogers 1974). This has led Nicholls (1975), and Sun and Hanson (1975) to suggest that such rocks are related not to Benioff zones, but rather, to crustal fracturing and extension. In addition, Arculus and Johnson (1978) concluded that the distribution of high-K_2O alkalic lavas and the depth of the Benioff zone showed little or no correlation within most island arcs.

Cooke (1966) compared his lavas with the leucite-bearing lavas of Italy's Roman Region. These rocks, like Cooke's rocks, have less than 1% TiO_2. Alvarez (1972) suggested that the Italian rocks were derived by partial fusion of argillaceous sediments in a subduction zone under the Roman Region, but Cundari (1979) discounted this process in favour of mantle doming along a graben structure similar to those found in Africa and elsewhere.

Except for alkalic lavas of the Absaroka Range, Wyoming (Nicholls and Carmichael 1969), most of the Cenozoic mafic alkalic lavas of the mid-western U.S.A. appear to be high-TiO_2 lavas (e.g. Leeman 1974; Noble and Parker 1974; Best and Brimhall 1974). Gest and McBirney (1979) suggested that both the Absaroka lavas and the Roman lavas were the result either of the partial melting of the garnet-bearing rocks, possibly eclogite, or the anatexis of lower crustal rocks. The authors further suggested that these rocks could not be related to Benioff zones. This would indicate that an island-arc environment would not be required for the development of the trachytic lavas in the Kirkland Lake area as proposed by Cooke (1966).

Extension tectonics and normal faulting are associated with the low-TiO_2 trachytes (shoshonites) in the Absaroka Range and elsewhere (Gest and McBirney 1979). Similar conclusions have been made by Cundari (1979) in the Roman Province and by Preto (1977) for the (low-TiO_2) alkalic lavas of the Nicola Group in British Columbia. In the Kirkland Lake area, the alkalic volcanic rocks appear to be closely associated with the Kirkland Lake - Larder Lake and Destor - Porcupine Fault Zones.

If crustal fracturing and extension are the main prerequisites required for low-TiO_2 alkalic volcanism, the tectonic environment required for these rocks differs little from that required for high-TiO_2 alkalic volcanism found in Africa. The magmas for low-TiO_2 alkalic rocks, however, appear to require the partial melting of eclogite or the anatexis of crustal rocks (Gest and McBirney 1979), possibly sediments (Alvarez 1972), and the magmas for high-TiO_2 alkalic rocks appear to require the partial melting of sources in the mantle (Baker *et al.* 1977). Melting of continental lower crustal rocks by heat from a rising mantle diapir would explain the generation of low-TiO_2 magmas. Mixing of magmas generated in the lower crust and in the mantle could also explain the spectrum of rocks along many rift zones, and the low abundance of trachytic rocks in ocean-island settings such as Hawaii. Hansen (1978) proposed that the trachytes of Ross Island were formed by a mixing of crustal rock and mantle derived magma.

In the Kirkland Lake area, the dominance of low-TiO_2 trachytic rocks would suggest that rifting of continental crust occurred above a heat source. The presence of high-TiO_2 alkali basalt and nephelinite along the Destor - Porcupine Fault Zone would suggest that a mantle diapir could have provided the heat required, as well as minor amounts of volcanic material.

Until Sr isotope and REE data are obtained on trachytes and more mafic alkalic rocks in the Kirkland Lake area or on comparable Archean rocks elsewhere the magma sources for these rocks can only be speculated upon. Judging from data on Cenozoic trachytes, the magmas for the Kirkland Lake trachytes may have been derived by the partial melting of eclogite or crustal rocks, possibly sediments, because of their low-TiO_2 contents. Some partial melting of the mantle may have also been involved.

Alkalic Intrusive Rocks

In the Kirkland Lake area, monzonite, quartz monzonite, syenite, and granodiorite stocks and dikes (K-rich granitic rocks) cut all other rock types except diabase. Most of these stocks are located either near or along the Destor - Porcupine and Kirkland Lake - Larder Lake Fault Zones, where alkalic volcanic rocks are known to occur. Elsewhere, where alkalic volcanic rocks and K-rich granitic rocks are known to occur close together, this close spatial association has been used to link the 2 groups of rocks (for example, Preto 1977; Smith and Longstaffe 1974).

In most Archean greenstone belts, K-rich granitic rocks do not appear to have been related to alkalic volcanism. However, these rocks generally form the youngest rocks in Archean terrains, and are considered to be distinctive from the Na-rich tonalites and trondhjemites (Anhaeusser *et al.* 1969, 1975b; Viljoen and Viljoen 1969a; Arth and Hansen 1975; Krupicka 1975; Glikson 1976a, 1979b; Drury 1979; Naqvi *et al.* 1978).

Up until the time writing, there has been little agreement on the origin of these K-rich granitic rocks. Many have favoured the reworking of primary granitic crust, or the remelting of tonalitic rocks to generate these rocks (Hunter 1974; Glikson 1976a, 1979b; Glikson and Lambert 1976; Bridgwater and Collerson 1976; Collerson and Fryer 1978; Fyfe 1978b). In many places, the K-rich granites intrude tonalitic rocks a considerable distance away from the nearest greenstone belt (Hunter 1973; Glikson 1979b).

Other authors have denied that there is a relationship between the older tonalitic rocks and the K-rich granitic rocks. Naqvi *et al.* (1978) suggested, instead, that K-rich granitic rocks were formed by the partial melting of the lower crust at greater depths than were the tonalitic rocks. Because Sr isotopes do not support the reworking of older granitic rock, Davies and Allsopp (1976) and Robb (1978) have proposed that K-rich rocks were formed from the mantle as the earth's crust thickened and cooled downward.

Rare-earth element data suggest that the magmas for quartz-bearing Archean K-rich granitic rocks may have been derived from sedimentary rocks, whereas quartz-deficient types may have been derived from a mixed source consisting of garnet-bearing peridotite and eclogite. Arth and Hanson (1975) concluded that the quartz monzonite of northeastern Minnesota represented a magma which was derived from the partial melting of a greywacke at crustal depths, leaving a residue of plagioclase, amphibole, garnet, and pyroxene or biotite. Glikson (1976a) also stated that the REE data for the Hood granite and for the Nelspruit migmatites of the Barberton area could be interpreted in terms of partial melting of greywackes, but he favoured the partial melting of earlier formed tonalitic rocks and some concomitant fusion of this magma with ultramafic rocks. Similarly, Drury (1979) has proposed that the K-rich granites of the Yellowknife area represented the partial melting of metasediments.

For the syenites and syenodiorites of northeastern Minnesota, Arth and Hansen (1975) favoured partial melting of a mixed garnet-bearing peridotite-eclogite source located in the mantle. In the Barberton area, this type of source was also favoured by Glikson (1976a) for the Bosmanskop pluton which had been referred to earlier as a syenite by Viljoen and Viljoen (1969a).

The limited Sr isotope and REE data for K-rich granitic rocks suggest that these rocks had sources of magma that were distinctive from those of the Na-granites. This data also suggests that they were not derived by the partial melting of Na-rich granitic rocks but rather that the magmas for the K-rich granites may have come from partly melted garnet-bearing sedimentary and mafic rocks as has been proposed for Cenozoic trachytic magmas by Gest and McBirney (1979), Alvarez (1972), and Hanson (1978). This would suggest that the K-rich granitic rocks and the alkalic volcanic rocks in the Kirkland Lake area could have had the same magma sources.

NOMENCLATURE AND CLASSIFICATION

The alkalic volcanic rocks have been named and classified according to Irvine and Baragar (1971), and the K-rich granitoid rocks have been named and classified according to Streckeisen (1976). Strict correspondence to the 2 classifications cannot be made in this report.

The chemical classification of Irvine and Baragar (1971) relies heavily on alkalies and silica, and tends to classify the mafic and ultramafic subalkalic volcanic rocks as alkalic rocks where they have been slightly enriched in alkalies, particularly near intrusions of K-rich granitic rocks. Careful field and petrographic studies and the consideration of other chemical parameters are required to confirm whether or not a rock is alkalic. In addition, trachytes with minor amounts of normative quartz, are classified as calc-alkalic andesites and dacites. Again, careful field and petrographic studies are required to distinguish them from other calc-alkalic rocks that are unassociated with alkalic volcanic rocks.

As discussed previously, the granitoid rocks have been grouped into Na-rich and K-rich rocks. Streckeisen's (1976) classification can be applied strictly to the Na-rich (plagioclase-rich) rocks which include quartz diorite, tonalite, and trondhjemite. Streckeisen's (1976) classification is a petrographic one which does not distinguish K-feldspar from albite, it does not require that syenites be K-rich rocks. To avoid this confusion, albite-rich granitic rocks are referred to as syenodiorites in this study.

FIELD AND PETROGRAPHIC DESCRIPTIONS OF THE ALKALIC VOLCANIC AND ASSOCIATED INTRUSIVE ROCKS

Alkalic Volcanic Rocks

Near Kirkland Lake, the alkalic rocks form well exposed rock ridges but elsewhere, particularly in the map area, these rocks tend to form only scattered, low-lying, poorly exposed knolls.

The strongest evidence for alkali basalt and nephelinite comes from 3 closely spaced diamond-drill holes in Harker Township put down by Anatole Resources Incorporated in 1973. One hole intersected altered komatiitic lavas, and the other 2 intersected a succession of lavas and tuffs with high alkali contents. The mafic lavas within the 2 latter holes contain more than 3% TiO_2, and have chemical compositions similar to alkali basalts and nephelinites (Table 4). In these rocks, blue-green amphibole and some exsolved ilmenite pseudomorphic after subophitic augite grains, 0.05 to 2 mm in maximum dimension, occur with euhedral to subhedral magnetite grains, 0.5 to 2 mm in size. The plagioclase grains are euhedral to subhedral, and are partly saussuritized. The less mafic but more alkali-rich varieties contain hexagonal pseudomorphs of feldspathoids (possibly leucite), 0.5 to 2 mm across, composed largely of acicular alkali feldspar and needle-shaped, blue-green actinolite. Euhedral to subhedral, blue-green hornblende grains, interlocked with grains of ilmenite and albite, form the rest of these rocks.

The tuffaceous varieties contain rock fragments of the above rock types as well as isolated grains of feldspar, ilmenite, hornblende, and pseudomorphic feldspathoids set in a matrix rich in chlorite and epidote.

Only altered iron-rich tholeiitic basalt could be mistaken for alkali basalt or nephelinite. However, iron-rich tholeiitic basalt that contains more than 3% TiO_2 tends to have about 18% FeO_T. The samples of alkali basalt and nephelinite just described have 12 to 14% FeO_T, and therefore have much higher TiO_2/FeO_T ratios. The ilmenite grains in these samples are unaltered and the presence of pseudomorphic feldspathoids in some of these rocks suggests that they are true alkali basalts and nephelinites.

The selected alkali basalts and nephelinites require further chemical study to determine their origin. The composition of the rocks suggests that they may have been derived by limited partial melting of the mantle (Green and Ringwood 1967a). If so, they may represent, instead of tholeiitic magma as postulated by

TABLE 4: CHEMICAL COMPOSITION OF ALKALIC VOLCANIC ROCKS.

Major Components in %

Sample Number	**1**	**2**	**3**	**4**	**5**
Field Number	**73J1837**	**73J1838**	**73J1839**	**73J1841**	**73J1836**
SiO_2	44.80	43.00	47.10	52.90	65.40
Al_2O_3	12.20	11.80	13.20	13.50	13.90
Fe_2O_3	12.80	14.20	14.10	13.60	3.85
MgO	4.53	4.83	3.86	2.94	1.87
CaO	10.30	10.00	8.00	6.03	5.00
Na_2O	5.50	5.18	6.10	6.69	4.78
K_2O	0.75	1.91	.97	.34	2.94
TiO_2	3.23	3.24	3.27	2.72	0.96
P_2O_3	-	-	-	-	-
MnO	0.27	0.24	0.21	0.20	0.09
L.O.I.**	6.25	5.27	4.40	1.45	2.15
Total:	100.6	100.1	100.2	100.4	100.9

Notes:
*Total Fe reported as Fe_2O_3
**Loss on ignition

Molecular Norms*

Sample Number	**1**	**2**	**3**	**4**	**5**
Ilmenite	4.816	4.843	4.762	3.878	1.354
Orthoclase	4.748	12.120	5.997	2.058	17.604
Albite	24.791	10.086	33.222	57.626	43.448
Anorthite	6.838	3.545	6.036	5.943	7.881
Magnetite	5.293	5.317	5.213	4.515	1.126
Hematite	-	-	-	-	0.985
Wollastonite	1.612	-	-	-	1.668
Quartz	-	-	-	-	15.480
Diopside	26.769	27.783	18.761	10.844	10.453
Forsterite	-	0.311	1.318	2.163	-
Fayalite	-	0.134	0.675	1.774	-
Nepheline	16.835	23.887	14.413	2.302	-
Hedenbergite	8.298	11.973	9.604	8.897	-

Notes:

1. Grey, aphanitic, magnetic nephelinite (Na-alk. bas. ser.).

2. Greenish-grey, fragmental, magnetic nephelinite (Na-alk. bas. ser.).

3. Grey, 0.5 mm grained, massive, magnetic nephelinite (Na-alk. bas. ser.).

4. Dark bluish-grey, massive, magnetic mugearite (Na-alk. bas. ser.)

5. Cherty reddish-brown, aphanitic, massive magnetic calc-alkalic dacite interlayered with alkalic basalts.

*Barth-Niggle Katanorm (Chayes and Metais 1963)

Arndt (1976, 1977), an early partial melt of a mantle diapir which further melted to produce komatiitic magmas. To date, no such rocks have been found associated with komatiites in other greenstone belts and Nesbitt *et al.* (1979), have argued against such a process.

Alkalic basalts and nephelinites are associated with tholeiitic basalts in oceanic island settings (Macdonald and Katsura 1964). Basalts with less than 1% TiO_2 are rare in such settings. In the Kinojevis Group, more than half of the basalts are magnesium-rich tholeiitic basalts with less than 1% TiO_2. In addition, if these rocks were part of the Kinojevis Group, they should be recognizable elsewhere in the area.

The remaining alternatives are, either that these alkali basalts and nephelinites are a phase of the alkalic volcanism which produced the trachytic lavas and tuffs, or, that they represent a separate phase of volcanism and their association with trachytic volcanic rocks along the Destor - Porcupine Fault Zone is only circumstantial. Until more can be learned about the field relationships between the alkali basalts and nephelinites and the other groups of rocks, and more detailed chemical analyses are done to determine their significance, the presence of these rocks can be used only in connection with other evidence concerning the development of the trachytic rocks in the Kirkland Lake area.

Less mafic alkalic volcanic rocks (trachybasalt, tristanite, trachyte) which occur in the map area are like those rocks near Kirkland Lake described in detail by Cooke (1966), Cooke and Moorhouse (1969), and Hyde (1978). An improvement in these extensive descriptions is beyond the scope of this report.

Alkalic Intrusive Rocks

Descriptions of syenodiorite, syenite, monzonite, quartz monzonite, granite, and granodiorite that outcrop near and along the Destor - Porcupine Fault have been given by Satterly (1949a, 1949b, 1952a, 1952b) and by Prest (1953, 1957). Near Kirkland Lake, these rock types have been described by Thomson (1943, 1949, 1950), Thomson and Griffis (1944), Abraham (1950), MacLean (1956), and Lawton (1959).

These rocks tend to form low-lying, massive, rarely gneissic, well-jointed outcrops. In thin section, the feldspars and mafic minerals are revealed to be well preserved. These rocks have 20 to 60% orthoclase or perthite and 0 to 25% quartz by which they can be classified. Hornblende or biotite are the main mafic mineral in these rocks; amounts vary from 10 to 50%. In some syenites and syenodiorites, nepheline is present as described by Lawton (1959). Where porphyritic, the phenocrysts can be plagioclase, orthoclase, or perthite. K-rich feldspar rims the plagioclase phenocrysts.

GEOLOGY OF LIGHTNING RIVER AND MAGUSI RIVER AREAS

STRATIGRAPHIC UNITS

In this report, data from the detailed mapping of the Lightning River and the Magusi River map areas are used to illustrate the development of the Abitibi Greenstone Belt in the Kirkland Lake area. These 2 adjoining map areas form an ideal type area for the greenstone belt, because they, together, encompass the northern limb of the synclinorium over a width of 35 km (Figure 19). The total volcanic succession in this area is 30 km in thickness and is similar to that found on the southern limb of the synclinorium (Table 5). The Destor - Porcupine Fault Zone transects the southern part of the Lightning River area. A cross section of the rocks in the 2 map areas is shown in Figure 20. The detailed geology of the Lightning River and Magusi River areas is shown on Maps P.2433 and P.2434, back pocket, at a scale 1:63 360.

The oldest rocks are calc-alkalic volcanic rocks. These rocks occur farther to the east, in Quebec, where they were named the "Hunter Mine Group" by Dimroth *et al.* (1973). Dimroth described the Hunter Mine Group as being composed of layered volcanic rocks and sedimentary rocks, consisting of cherty tuff, tuff-breccia, argillite, and banded iron formation. These rocks extend westwards into Roquemaure Township, Quebec, where they were described by

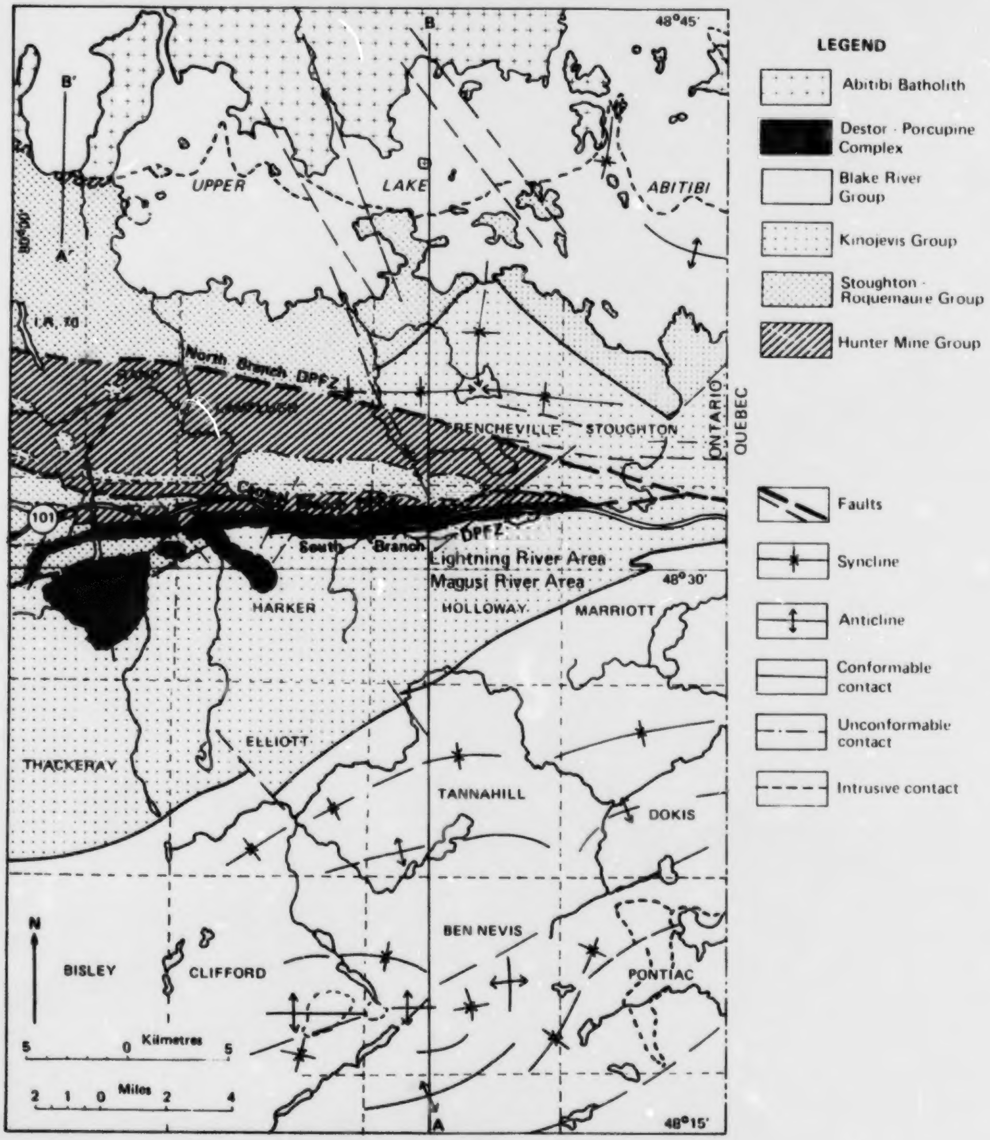


Figure 19. Geological map of the Magusi River and Lightning River areas.

Eakins (1972), as mainly composed of fragmental and hypabyssal andesites, dacites, and rhyolites with minor cherty tuff and iron formation. Eakins (1972), described the calc-alkalic rocks as being overlain by a succession of ultramafic to mafic lavas called the Stoughton - Roquemaure Group by Jensen (1976a, 1978b). Rocks of the Hunter Mine Group occur below the Stoughton - Roquemaure Group at the western end of Lake Abitibi, in the southern parts of Frecheville, Lamplugh and Rand Townships, Indian Reserve 70, and in the northern parts of Holloway, Harker and Garrison Townships (Figure 19). At the western end of Lake Abitibi and in Rand Townships, they are mainly fragmental and hypabyssal andesites and dacites. In Frecheville, Lamplugh and Holloway Townships, however, they are mainly pillowed lavas consisting of calc-alkalic basalts and andesites which are interlayered by cherty tuffs and tuff-breccias of andesite, dacite, and rhyolite. In Harker and Garrison Townships, the Hunter Mine Group grades into cherty tuff and tuff breccia interlayered by banded iron formation and argillite. These rocks could be correlated with the sedimentary rocks farther to the west in the southern parts of Munro Township (Satterly 1952a), and are referred to as the Porcupine Group by Pyke (1980). The lateral facies changes in the Hunter Mine Group suggest that a volcanic pile formed in the vicinity of the Lake Abitibi Batholith, and that the flanks of this volcanic pile sloped downward to the east, south, and west into a basinal area with sedimentation.

Rocks of the Stoughton - Roquemaure Group unconformably overlie the Hunter Mine Group. The Stoughton - Roquemaure Group consists of peridotitic and basaltic komatiites and MgO-rich tholeiitic basalt, with minor FeO-rich tholeiitic basalt at its base and top. It is exposed as a thick monoclinal volcanic succession along the southern shore and adjacent islands of Lake Abitibi, and has a thickness of up to 10 600 m (*see* Figure 6). Dunite, peridotite, and gabbro are associated with the basal units of the group. These basal units occur farther south in Holloway, Harker and Garrison Townships where they are thicker than in the north. This suggests that volcanism for the Stoughton - Roquemaure Group was more extensive to the south.

The Kinojevis Group lies above the Stoughton - Roquemaure Group, and is about 10 500 m in thickness. It consists of alternating map units of magnesium-rich and iron-rich tholeiitic basalt which have minor amounts of tholeiitic andesite, dacite, and rhyolite present toward the top. The Kinojevis Group occurs north and south of the Destor - Porcupine Fault (*see* Figure 19).

The Kinojevis Group is overlain by the Blake River Group which is approximately 10 000 m in thickness. The Blake River Group consists mainly of calc-alkalic andesite and basalt with some calc-alkalic dacite and rhyolite concentrated mainly toward its top. It is cut by numerous stocks of gabbro, quartz gabbro, hornblende gabbro, and diorite. Two rhyolitic domes cut by quartz diorite stocks and surrounded by massive dacite and rhyolite flow-breccia, occur in Pontiac and Clifford Townships (Jensen 1975a, 1975b). They are thought to represent centres of volcanism for most of the calc-alkalic rocks in the area. Thick lenses of poorly bedded volcaniclastic rock, thought to be slump deposits off the volcanic centres,

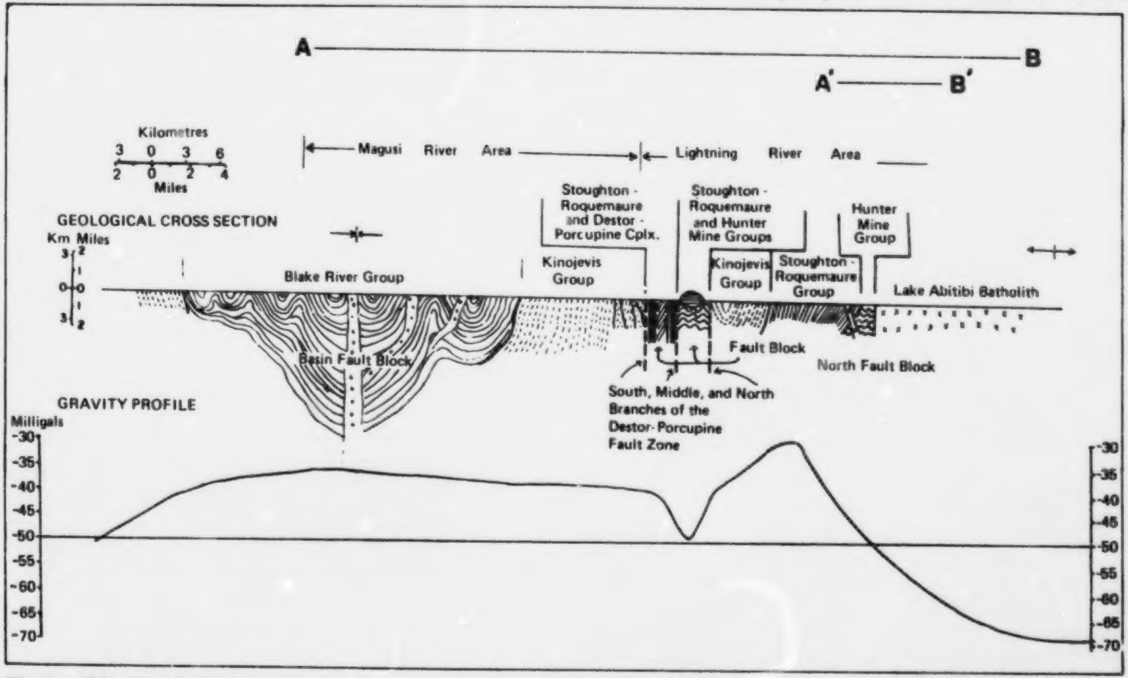


Figure 20. Geological cross-section and gravity profile of the Magusi River and Lightning River areas. Locations of A-B, A'-B' are shown on Figures 19 and 22.

TABLE 5: STRATIGRAPHIC SUCCESSION OF THE KIRKLAND LAKE AREA.

SOUTH LIMB OF SYNCLINORIUM
Upper Supergroup
- Timiskaming Group
  - Volcanic rocks: Mafic, intermediate, felsic trachyte, and K-rich dacite and rhyolite flows and tuffs.
  - Sedimentary rocks: Fluvial conglomerate, sandstone, and argillite.
  - Intrusions: Stocks and dikes of syenodiorite, syenite, quartz monzonite, and lamprophyre.
- Blake River Group
  - Volcanic rocks: Calc-alkalic basalt, andesite dacite and rhyolite flows and tuffs.
  - Sedimentary rocks: Volcaniclastic slump deposits.
  - Intrusion: Stocks and dikes of gabbro, quartz gabbro, hornblende gabbro, diorite, quartz diorite, and subvolcanic rhyolite domes.
- Kinojevis Group
  - Volcanic rocks: Mg-rich and Fe-rich tholeiitic basalts, and tholeiitic andesite, dacite and rhyolite flows and tuffs.
  - Sedimentary rocks: Thin interflow argillite and chert.
  - Intrusions: Sills of Mg-rich and Fe-rich gabbro.
- Larder Lake Group
  - Volcanic rocks: Flows of peridotitic and basaltic komatiite, and Mg-rich tholeiitic basalt, and minor Fe-rich tholeiitic basalt, and minor Fe-rich tholeiitic basalts and interflow rhyolite tuff- breccias.
  - Sedimentary rocks: Turbiditic conglomerate, greywacke and argillite, and iron formation chert, limestone, and dolostone.
  - Intrusions: Sills and stocks of peridotite, pyroxenite, and gabbro.

Lower Supergroup
- (Unnamed Unit)
  - (Conglomerate with trachyte and syenodiorite pebbles).
- Skead Group
  - Volcanic rocks: Mainly calc-alkalic rhyolite tuff-breccia with some calc-alkalic basalt, andesite and dacite flows and tuff- breccias.
  - Sedimentary rocks: Chert and cherty argillite.
  - Intrusions: Stocks of feldspar porphyry and quartz diorite.
- Catherine Group
  - Volcanic rocks: Mg-rich and Fe-rich tholeiitic basalts.
  - Sedimentary rocks: Interflow chert.
  - Intrusions: (None mapped)
- Wabewawa Group
  - Volcanic rocks: Peridotitic and basaltic komatiite, Mg-rich tholeiite basalt and minor Fe-rich tholeiitic basalt, and a few interflow rhyolite tuffs.
  - Sedimentary rocks: (None mapped)
  - Intrusions: Layer sills (possibly flows) of dunite, pyroxenite, and gabbro.
- Pacaud Tuffs (Ridler 1970)
  - Volcanic rocks: Calc-alkalic andesite, dacite, and rhyolite tuffs.
  - Sedimentary rocks: Chert, argillite, iron formation.
  - Intrusions: Trondhjemite of Round Lake Batholith.

occur interlayered with the volcanic flows in Tannahill, Dokis, and Ben Nevis Townships.

The Destor - Porcupine Complex occurs between 2 branch faults of the Destor - Porcupine Fault Zone (*see* Figure 19). It is composed of sedimentary rocks, alkalic volcanic rocks, and alkalic to subalkalic intrusive rocks.Conglomerate containing pebbles of alkalic intrusive and extrusive rock along with greywacke, argillite, sandstone, and iron formation form the sedimentary rocks. The sedimentary rocks are interlayered by alkalic volcanic rocks composed of alkali basalt, hawaiite, mugearite, Na-trachyte, benmorite, and alkali-rich dacite and rhyolite, and are cut by albite syenodiorite dikes and stocks. A few of the larger stocks are composed of granodiorite and quartz monzonite, and a few of the alkalic to subalkalic intrusions cut the Kinojevis and Blake River Groups to the south as well.

To the north, the Lake Abitibi Batholith intrudes the Hunter Mine Group and the lower parts of the Stoughton - Roquemaure Group. The batholith is a composite intrusion 50 by 30 km across, composed mainly of massive and gneissic trondhjemite.

TABLE 5. CONTINUED

NORTH LIMB OF SYNCLINORIUM

Upper Supergroup

- Destor - Porcupine Complex
  - Volcanic rocks: Alkali basalt, benimorite, hawaiite, mugearite, and alkali-rich dacite and rhyolite.
  - Sedimentary rocks: Turbiditic conglomerates, greywacke, argillite, and ironstone.
  - Intrusions: Stocks and dikes of syenodiorite, granodiorite, and quartz monzonite.
- Blake River Group
  - (*See* south limb)
- Kinojevis Group
  - (*See* south limb)
- Stoughton - Roquemaure Group
  - Volcanic rocks: Flows of peridotitic and basaltic komatiite and Mg-rich tholeiitic basalt and minor Fe-rich tholeiitic basalt and minor interflow calc-alkalic rhyolite tuff and cherty tuff.
  - Sedimentary rocks: Minor chert and iron formation.
  - Intrusions: Sills and stocks of peridotite, pyroxenite, and gabbro.

Lower Supergroup

- Hunter Mine Group
  - Volcanic rocks: Mainly calc-alkalic dacite and andesite tuff- breccia with some calc-alkalic basalt, andesite and dacite flows.
  - Sedimentary rocks: Cherts, iron formation and turbiditic greywacke, and argillite (may be equivalent to Porcupine Group (Pyke 1979))
  - Intrusions: Dikes of quartz-feldspar porphyry and trondhjemite of Lake Abitibi Batholith.

AEROMAGNETIC DATA

The general geology of the Kirkland Lake Area is reflected by aeromagnetic maps 7084 and 7085 (Ontario Department of Mines - Geological Survey of Canada 1970a, 1970b). For detailed geological study, some individual units can be recognized and extrapolated on the aeromagnetic maps, but, in general, the maps are most useful for outlining groups of rocks. Magnetic susceptibility maps are being generated at the University of Toronto in the hope of improving the sensitivity of the present data (Letros *et al.* 1983).

In the Lightning River and Magusi River areas, the calcalkalic rocks of the Hunter Mine Group and Blake River Group have weak magnetic expressions because of their low magnetic mineral contents (Figure 21). The isolated magnetic anomalies in the Blake River Group are the result of gabbroic intrusion and metamorphic aureoles around felsic intrusions. Weak magnetic expressions also occur over the granitic rocks of the Lake Abitibi Batholith.

The high magnetic anomalies are caused by iron-rich tholeiitic basalts, tholeiitic andesites, and altered ultramafic rocks rich in magnetite. In the northern part of the Magusi River area, the Kinojevis Group has a distinctively striped pattern caused by groups of iron-rich tholeiitic basalt lava flows alternating with groups of magnesium-rich tholeiitic basalt having a lower magnetite content than the iron-rich tholeiitic basalt. This striped pattern intensifies upward in the group because of the iron-enrichment trend in the lavas. At the top, there is a steep enough gradation from the high magnetics of the Kinojevis Group into the low magnetics of the Blake River Group to suggest that the Kinojevis Group dips under the Blake River Group. All the lava flows in this area dip and face south.

The Stoughton - Roquemaure Group shows a less intense magnetic pattern than the Kinojevis Group because of its abundant basaltic komatiites and Mg-rich tholeiitic basalts of low magnetite content. The zones of high magnetic expression in the Stoughton - Roquemaure Group are the result of iron-rich gabbroic rocks and serpentinized ultramafic rocks. Conspicuous examples are the serpentinized peridotites and iron-rich gabbroic rocks which occur with the Hunter Mine Group in the southern part of the Lightning River area.

The 3 main branches representing the Destor - Porcupine Fault Zone can be recognized by the abrupt changes which occur in the magnetic patterns across these faults. For example, the striped pattern of the Kinojevis Group is truncated by the south fault of the Destor - Porcupine Fault Zone (Figure 21). The Kinojevis Group is juxtaposed against a variety of rock types that includes sedimentary rocks, alkalic volcanic and intrusive rocks, peridotites, and komatiites that cause an irregular magnetic pattern on the north side of the fault. Along the north fault of the Destor - Porcupine Fault Zone, the strong magnetic patterns of the Stoughton - Roquemaure and Kinojevis Groups on the north side are truncated against the weak magnetic pattern on the south side caused by the Hunter Mine Group.

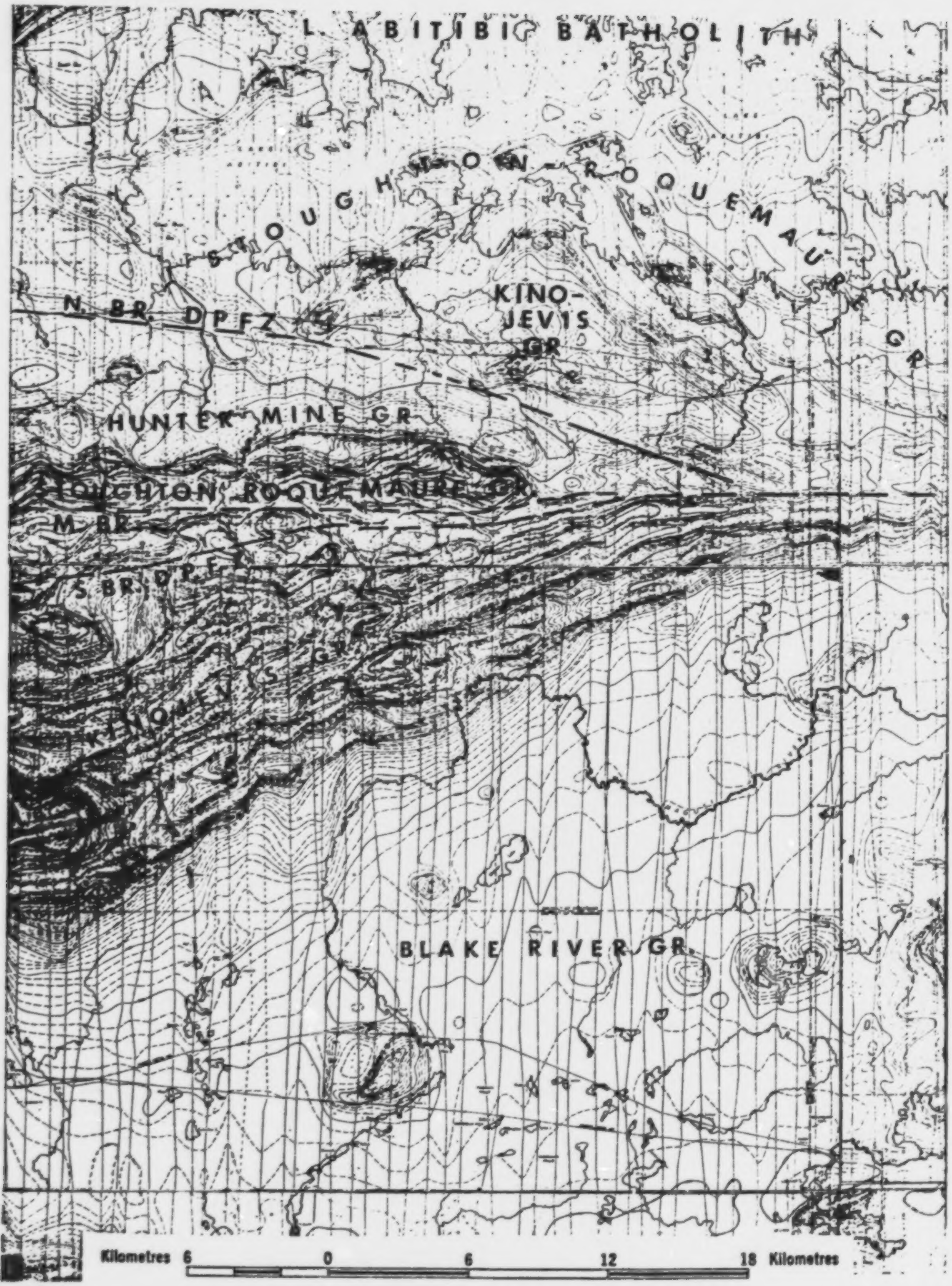
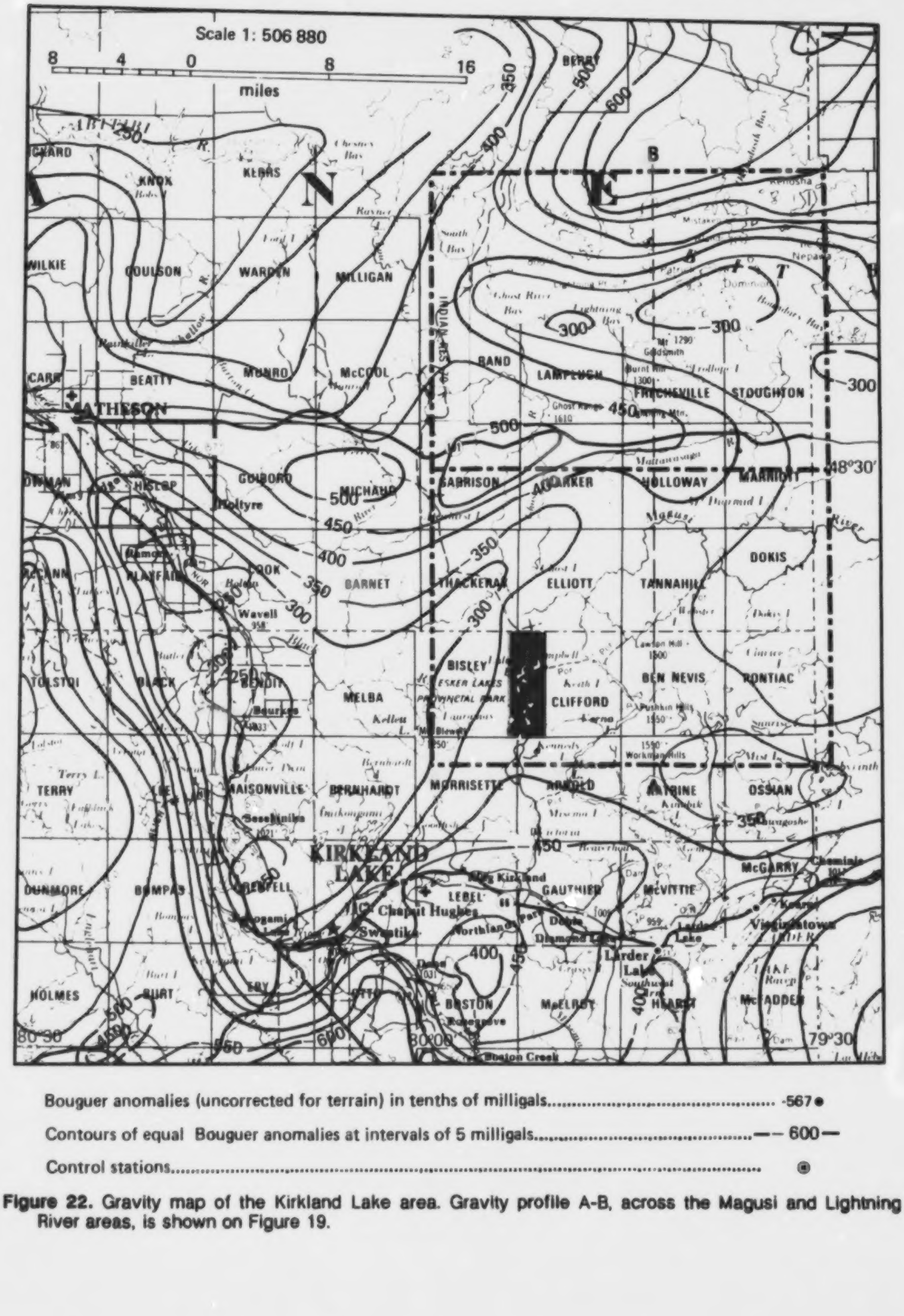


Figure 21. Aeromagnetic map of the Lightning River and Magusi River areas.



Bouguer anomalies (uncorrected for terrain) in tenths of milligals... -567●

Contours of equal Bouguer anomalies at intervals of 5 milligals..——600—

Control stations... ⊙

Figure 22. Gravity map of the Kirkland Lake area. Gravity profile A-B, across the Magusi and Lightning River areas, is shown on Figure 19.

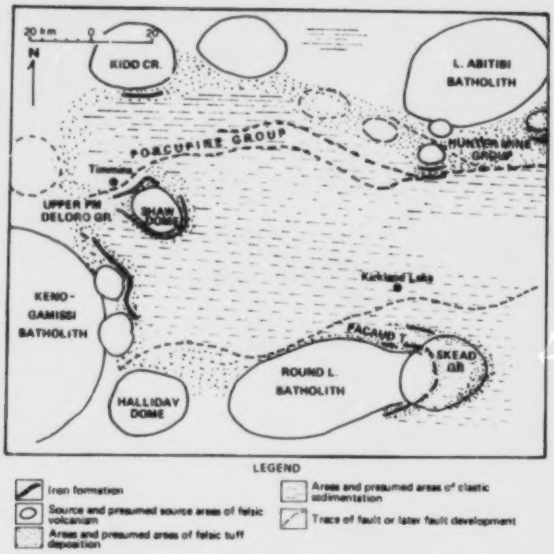


Figure 23. Map diagram of the calc-alkalic volcanism and clastic sedimentation of the lower supergroups. Note that the Skead Group appears to have been a volcanic pile which developed on the edge of the volcanic pile represented by the Pacaud Tuffs.

ODM-GSC Maps 7084G and 7085G (Ontario Department of Mines - Geological Survey of Canada 1970a, 1970b) show that the distinctive striped pattern of the Kinojevis Group extends west around the nose of the synclinorium and encloses the weaker magnetic pattern of the Blake River Group. On the southern limb of the synclinorium, the striped pattern of the Kinojevis Group is interrupted by numerous felsic intrusions. This is also apparent farther to the west, southeast of Timmins, where the pattern is interrupted by diabase dikes and felsic intrusions.

The Kirkland Lake - Larder Lake Fault Zone can be recognized on Maps 7084G and 7085G (Ontario Department of Mines - Geological Survey of Canada 1970a, 1970b) by the abrupt change in direction of the magnetic contours. The irregular patterns south of the fault zone and in the fault zone are caused by komatiitic, tholeiitic, calc-alkalic, and alkalic volcanic rocks and sedimentary rocks with iron formation, all these rocks are folded and intruded by plutons ranging in composition from peridotite to felsic plutons.

GRAVITY DATA

The available gravity data (Figure 22) are based on widely spaced points of observation which greatly limits the interpretive value of such data. The contours of equal Bouguer anomalies in tenths of milligals, are based on the International Formula of 1930 (Gravity Map Series Number 21 and Number 80; Dominion Observatories, Ottawa 1968a, 1968b). The contours range from -65 milligals for granitic batholiths to -25 milligals for volcanic rocks. If the volcanic rocks are considered to have a specific gravity of about 2.8, and are assumed to grade into granitic rocks at depth, the volcanic rocks could be expected to extend to a depth of 7 to 10 km. However, such values should not be accepted until further gravity studies are made of the area.

The contours of equal Bouguer anomalies are consistent with the general geology of the area. The Lake Abitibi and Watabeag Batholiths have Bouguer anomalies of -65 to -60 milligals. Toward the central parts of the synclinorium, the Bouguer anomalies increase from -35 to -25 milligals, and suggest that the volcanic rocks extend much deeper in this part of the area. On the southern shore of Lake Abitibi, the contours increase to -30 milligals and closely reflect the distribution of the denser, steeply dipping succession of komatiitic and tholeiitic volcanic rocks of the Stoughton - Roquemaure and Kinojevis Groups. Along the Quebec - Ontario boundary, these contours extend south into the central parts of the synclinorium.

The intermediate contours of -50 to -40 that extend west from Lightning Mountain north of Highway 101, correspond to less dense felsic to intermediate calc-alkalic volcanic rocks of the Hunter Mine Group. The contours suggest that peridotitic and gabbroic rocks of the Stoughton - Roquemaure Group in this area do not extend very deeply. The intermediate contours south of the highway near Perry Lake would be a reflection of the steeply dipping sedimentary rocks of the Porcupine Group mapped by Satterly (1949a). Elsewhere, south of the highway, the contours are -35 to -25 milligals. These contours correspond to the mafic rocks of the Kinojevis Group and to the less dense overlying calc-alkaline volcanic rocks of the Blake River Group.

Along Highway 66, which approximates the location of the Kirkland Lake - Larder Lake Fault Zone, the Bouguer anomalies decrease from -40 to -60 milligals toward the south. This suggests that, although ultramafic and mafic volcanic rocks are abundant in this area, they are underlain by the less dense felsic volcanic and plutonic rocks observed farther to the south.

CALC–ALKALIC ROCK GROUPS OF THE LOWER SUPERGROUPS

Four known groups of calc-alkalic volcanic rock are located on the outer parts of the synclinorium, these are the: Hunter Mine Group, the Skead Group, the Upper Formation of the Deloro Group and the Pacaud Tuffs (*see* Figures 5 and 23, Table 5). These groups of calc-alkalic volcanic rock are largely felsic volcanic rocks which probably represented late-stage volcanism of developed volcanic piles (Goodwin 1968). The Skead Group, for example, overlies groups of tholeiitic and komatiitic volcanic rock (Ridler 1969, 1977).

Except for the Pacaud Tuffs which are older (*see* Figure 5, and Table 5), the Hunter Mine Group, the Upper Formation of the Deloro Group, and the Skead Group are overlain by a common group of komatiitic volcanic rocks that represent the base of the younger volcanic cycle (Upper Supergroup) preserved in the synclinorium (Jensen and Pyke 1980). If the komatiitic lavas were to have formed a common group overlying the separate groups of calc-alkalic

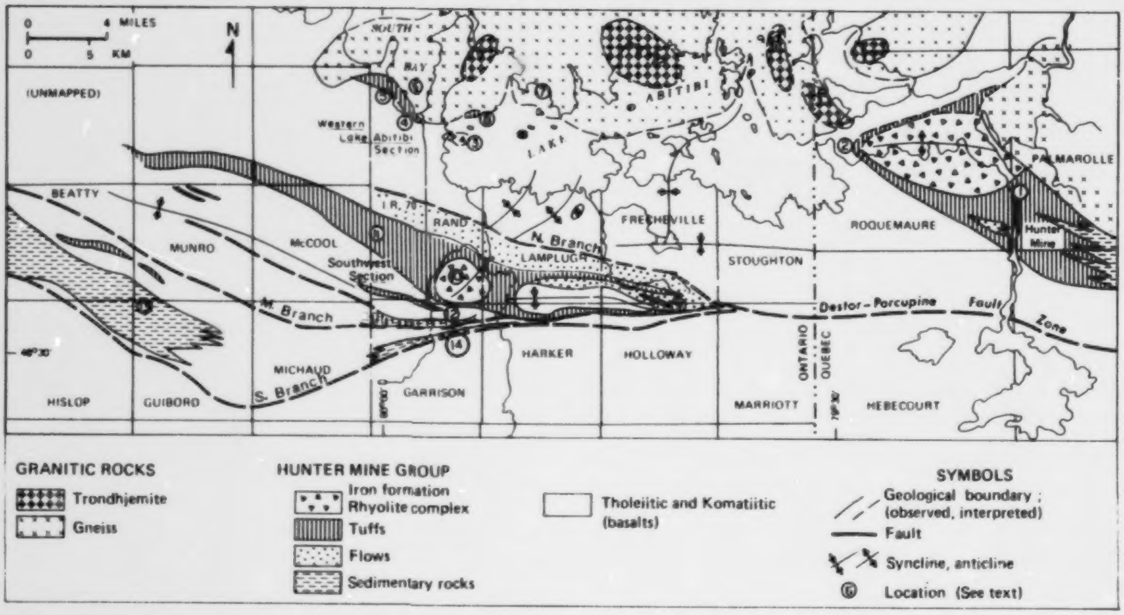


Figure 24. Map showing the distribution of rock types belonging to the Hunter Mine Group.

rock, then the komatiitic lavas must have been deposited in a basin among earlier, separately developed volcanic piles. The alternative view is that all the groups of calc-alkalic volcanic rocks formed 1 continuous group of rocks which became segmented as the komatiitic lavas were being deposited on them. As previously discussed, not all these groups of calc-alkalic rocks have the same age, nor is it possible, as will be shown, to fit them together physically.

In order that these different groups of calc-alkalic rock be overlain by 1 group of komatiites requires either 1 of 2 conditions, these are: 1. that the older volcanic piles be low enough or sink low enough to be flooded by one continuous sheet of komatiitic lavas, 2. that only the calc-alkalic volcanic material deposited toward the base of the volcanic piles at the edges of the basin be covered by komatiitic lavas.

The following study of the Hunter Mine Group will show that the latter supposition is preferred. The other groups of calc-alkalic volcanic rocks have enough features which are comparable to those of the Hunter Mine Group to suggest that they too were deposited at the interface between a basin and a volcanic pile.

HUNTER MINE GROUP

Distribution

The distribution of the rocks belonging to the Hunter Mine Group are shown in Figure 24. The group extends into Quebec where Dimroth *et al.* (1973) applied the same name 'Hunter Mine Series' to a sequence of 'mainly felsic flow breccias with minor agglomerate, rhyolite tuff, and greywacke' at the Hunter Mine in Palmarolle Township, Quebec, 19 km east of Ontario. From here, the Hunter Mine Group extends in a northwesterly direction into Roquemaure Township, Quebec, where it was considered by Eakins (1972) to be a "rhyolite complex".

In Ontario, the Hunter Mine Group is overlain by younger komatiitic and tholeiitic volcanic rocks and is exposed in the Lightning River map area at the western end of Upper Lake Abitibi and the southern shore of Lower Lake Abitibi. Here, it is referred to as the 'western Lake Abitibi section' of the Hunter Mine Group.

The Hunter Mine Group also occurs south of Lake Abitibi in the southern parts of Frecheville, Lamplugh, and Rand Townships, Indian Reserve 70, and in the northern parts of Holloway, Harker, and Garrison Townships. In these area it forms an area about 29 km long, up to 8 km wide, and is referred to as the "southwestern section" of the Hunter Mine Group. A Rb-Sr date of 2710 ± 2 Ma was obtained from a sample taken in Rand Township (Nunes and Jensen 1980).

West of the Lightning River area a group of sedimentary rocks is exposed which has been described by Satterly and Armstrong (1949) and Prest (1953) as underlying the mafic volcanic rocks of the area. These rocks occupy the same stratigraphic position as the Hunter Mine Group, and may extend east into the northern part of Garrison Township (Satterly 1953, p.10). The rocks of Garrison Township and nearby townships will be discussed under 'sedimentary rocks'.

Palmarolle Township, Quebec

Dimroth *et al.* (1973) described only the rocks they observed at the Hunter Mine (Figure 24). Along with the volcanic rocks and greywacke, these authors mentioned that a unit of jasper iron formation 6 m thick, occurred with the felsic volcanic rocks. Gelinas *et al.*

(1977a, 1977b) further described the Hunter Mine "Unit" as being mainly metarhyolite with minor meta-andesite and metabasalt. These authors did not provide chemical data to show this in their chemical study of the rocks in the area. From Palmarolle Township, the group extends northwest into Roquemaure Township, Quebec (*see* Figure 29, Location 1; Grenier 1967).

Goodwin and Shklanka (1967) found that iron formation associated with felsic volcanic rock represents a transition from a volcanic to a sedimentary association. However, Dimroth (1975b) concluded that iron formation could not be used as an indicator of bathymetry, but it could be used to infer that depths once existed below storm wave base (Dimroth 1977; Walker 1978), and that a fairly deep quiet water environment existed (Kimberley and Dimroth 1976). The Hunter Mine Group in Palmarolle Township, therefore was probably deposited in a quiet, possibly deepwater environment where periodic explosive felsic volcanism occurred not far distant from the basin of deposition, and that this section of the group represented the interface between a volcanic pile and a basin.

Roquemaure Township, Quebec

In Roquemaure Township, Eakins (1972) described the Hunter Mine Group as a 'rhyolitic complex' composed of layered fragmental rhyolite breccia interlayered with cherty tuff, chert, and iron formation cut by numerous penecontemporaneous rhyolitic porphyry bodies, and quartz-feldspar and feldspar dikes and sills. Lava flows and breccias composed of andesite occur interlayered with the rhyolitic rocks in a few places.

A diagnostic feature of the Hunter Mine Group was pointed out by Eakins (1972). He described some felsic rocks as containing up to 40% white feldspar phenocrysts, 0.2 to 0.5 cm in size, set in a grey to olive-green, fine-grained matrix and other felsic rocks as having quartz phenocrysts 0.5 to 1.5 mm in size.

Eakins (1972) also described the 'rhyolite complex' as forming the core of a domal anticline, the nose of which dipped west (Figure 24, Location 2). Eastward, the centre of the anticline is intruded by granitic rock so that the base of the Hunter Mine Group cannot be observed. From Eakins' map, only a minimum thickness of 3000 m can be inferred for the Hunter Mine Group in Roquemaure Township. Westward, northward, and southward, the Hunter Mine Group dips under ultramafic to mafic volcanic rocks which Eakins (1972) suggested had engulfed the felsic volcanic rocks.

Eakins (1972) noted that rocks of the Hunter Mine Group contain a schistosity which does not extend into the overlying mafic and ultramafic lavas. This feature of the group will be discussed elsewhere.

The chemical analyses from the Hunter Mine Group provided by Eakins (1972) are shown on Table 6. These chemical analyses, along with other chemical data from the group, have been projected on discriminant diagrams (Figures 25, 26, and 27). The 2 "andesites" of Eakins (1972) project as calc-alkalic basalts, and the rhyolites and dacites project as tholeiitic andesite dacite and rhyolite because of their high iron content (Figure 27). S, H_2O and CO_2 were not analyzed, so it is uncertain what proportions of the iron were present as sulphide minerals, oxides, carbonates, and silicates. The chemical data suggest that the dacites and rhyolites may have been calc-alkalic volcanic tuffs and breccias that either contained minor amounts of sulphide minerals or had minor amounts of iron oxide deposited with them. Eakins (1972) described the rhyolite breccias as having red chert fragments and the tuffs as containing grains of magnetite.

The rhyolite and dacite dikes, according to Eakins' map, are concentrated in the core of the anticline to suggest that this was an area of subaqueous explosive volcanism and that a dome resulted. During the formation of the dome and the volcanic activity, the previously deposited tuffs and sediments, including iron formation, were disrupted. This area of volcanism may have provided some of the volcanic material found in Palmarolle Township.

Western Lake Abitibi Section

Rocks of the Hunter Mine Group outcrop in a number of locations toward the west end of Lake Abitibi (Figure 24, Location 3 to 8). At Location 3, rocks of the group outcrop on several small islands that occupy an elliptical area 0.8 by 1.6 km. This area is surrounded by islands which contain exposures of only mafic volcanic rock. Here, the outcrops of the group are mainly calc-alkalic andesite, dacite, and rhyolite breccia and tuff-breccia cut by numerous dikes of calc-alkalic andesite, dacite, and rhyolite composition. Many of the breccia fragments, tuffs, and dikes contain up to 40% milky white plagioclase phenocrysts, 0.2 to 1 cm in size and quartz phenocrysts, 0.5 to 1.5 mm in size. The rocks closely fit Eakins' (1972) description of the rocks of the rhyolite complex in Roquemaure Township. The rocks also have schistosities which do not penetrate into the nearby mafic volcanic rocks. Unlike the rocks of Eakins (1972), these rocks contain no evidence of iron formation or argillite having been deposited with them.

On the southeastern shore of South Bay on Lower Lake Abitibi, the Hunter Mine Group is limited to 1 exposure of schistose dacite crystal tuff and tuff-breccia overlain by non schistose mafic pillowed lava tilted to the south (Figure 24, Location 4). This outcrop is at the top of a sequence of tuff and tuff-breccia, lying to the north that is thought to be about 650 m thick. An additional outcrop of strongly altered tuff-breccia occurs on the western shore of South Bay to indicate the sequence strikes northwest (Figure 24, Location 5). To the north and the west, at locations 6 and 7 (Figure 24), inclusions of partly assimilated schistose rhyolite tuff and tuff-breccia, cut by dikes and sills of dacite and rhyolite, quartz-feldspar porphyry, and feldspar porphyry identical to those at Location 3, occur in island and shoreline outcrops of gneissic tonalite and trondhjemite, respectively. The inclusions are up to 10 m in size, and suggest that the Hunter Mine Group occupied at least part of the area presently occupied by the Lake Abitibi Batholith.

One island with outcrops of bedded dacite tuff occurs near the margin of the Lake Abitibi Batholith; this suggest that in the western part of Upper Lake

TABLE 6: SEMI-QUANTITATIVE CHEMICAL ANALYSES OF ROCKS FROM THE HUNTER MINE GROUP IN ROQUEMAURE TOWNSHIP, QUEBEC (AFTER EAKINS 1972).

SAMPLE NUMBER	1	2	3	15	16	17
SiO_2	71.89%	68.85%	61.32%	56.45%	54.54%	73.62%
TiO_2	0.45	0.49	0.95	1.10	0.82	0.12
Al_2O_3	10.65	13.36	15.42	15.27	15.04	10.32
CaO	0.37	2.51	3.07	3.98	4.49	2.80
MgO	2.17	1.10	2.51	4.25	5.97	0.43
FeO*	9.71	5.73	8.10	8.61*	7.34*	3.65
Na_2O	0.07	3.50	5.85	2.93	4.30	4.20
K_2O	1.60	1.82	0.35	1.49	0.58	0.91
V	0.008	0.005	0.008	0.015	0.014	0.005
Cr	0.007	0.002	0.005	0.006	0.013	0.009
Fe (total)	7.55	4.44	6.28	6.75	5.69	2.83
Ni	19 ppm	17 ppm	31 ppm	116 ppm	82 ppm	10 ppm
Zn	48 ppm	75 ppm	140 ppm	88 ppm	53 ppm	71 ppm
Cu	37 ppm	32 ppm	64 ppm	77 ppm	126 ppm	55 ppm
Pb	8 ppm	12 ppm	12 ppm	12 ppm	38 ppm	8 ppm

Notes:
*Original determination was for total Fe. For purposes of calculation and comparison, total Fe has been converted entirely to FeO.

1. Rhyolite tuff from a road cut exposure on the southern side of the range road in lot 26, range V, immediately north of range IV-V line of Roquemaure Township (Analysis by the Quebec Department of Natural Resources (Q.D.N.R.)) Laboratory Branch, Q.D.N.R.).
2. Schistose rhyodacite flow breccia from a clean shoreline outcrop under a low clay cliff on the western side of a small peninsula on the southern side of Lake Abitibi in lot 31, range VI, close to the centre point of Roquemaure Township (Analysis by the Laboratory Branch, Q.D.N.R.).
3. Dacite flow breccia from a shoreline exposure in the southeastern corner of a small bay on the southern side of Lake Abitibi in the middle of lot 15, range V, of Roquemaure Township (Analysis by the Laboratory Branch, Q.D.N.R.).
15. Altered andesite lava; lot 48, range I, asbestos zone (Analysis by the Laboratory Branch, Q.D.N.R.).
16. Highly altered andesite lava; lot 38, range II (Analysis by the Laboratory Branch, Q.D.N.R.).
17. Rhyolite porphyry or feldspar-quartz porphyry from northern end of large outcrop 100 feet south of the range road in the southern end of lot 46, range V, Roquemaure Township.

Abitibi, additional portions of the Hunter Mine Group may be present (Figure 24, Location 8).

Major element analyses were done on 16 rock specimens from the western Lake Abitibi section. All are subalkalic (Figure 25), and range in composition from calc-alkalic basalt to calc-alkalic rhyolite, except for 1 (sample 436) which projects as a magnesium-rich tholeiitic basalt (Figure 25). They tend to be orthoclase-poor rocks (Figure 27). The samples are listed in Appendix 1, Tables 1 and 2, and their locations are shown in Map P.2433, back pocket.

The lack of iron formation and greywacke in the western Lake Abitibi section suggests that the Hunter Mine Group in this area was deposited in a more active volcanic environment and at a shallower water depth than were the rocks of Palmarolle and Roquemaure Townships.

Southwestern Section

The Hunter Mine Group is preserved in a large fault block located between the north and middle branches of the Destor - Porcupine Fault Zone. The fault block is wedge shaped and widens to the west (Figure 24). Outcrop is sparse in the northern part of the fault block through Rand, Lamplugh, and Frecheville Townships. Along the southern part of the block, the Hunter Mine Group is partly overlain by mafic and ultramafic rock, is severely sliced by faulting in many places, and is poorly exposed. This makes the units difficult to follow.

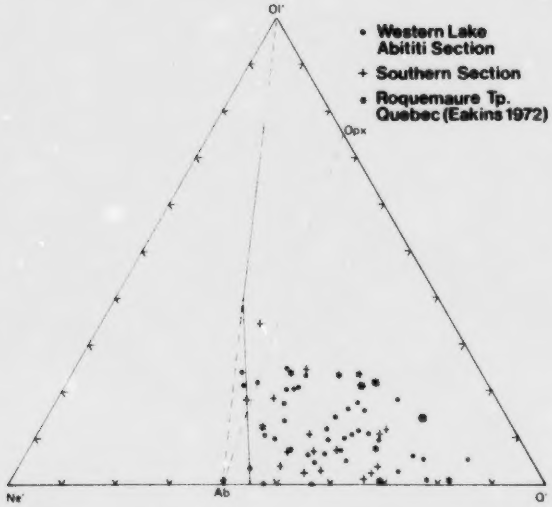


Figure 25. Ol'-Ne'-Q' projections of analyses from Hunter Mine Group (after Eakins 1972). Note that the black circle with a concentric circle indicates 2 plots from the western Lake Abitibi section which plotted at the same position.

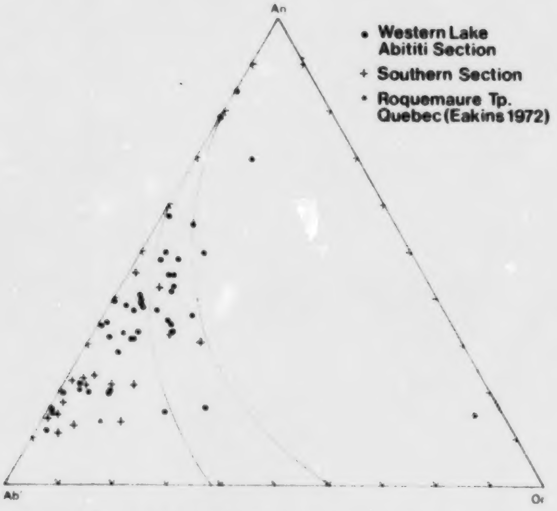


Figure 26. Ab'-An-Or projections of analyses from the Hunter Mine Group (after Eakins 1972).

In the northern part of the fault block, the scattered outcrops are calc-alkalic andesite, dacite, and rhyolite tuffs and tuff-breccias interlayered with amygdaloidal, massive to pillowed calc-alkalic basalt and andesite. A steeply dipping schistosity, subparallel to the dip and strike, is present in the rocks. The rocks are lower to middle greenschist metamorphic rocks. In the southern part of Frecheville Township, these rocks are folded into a syncline whose axis trends west (Figure 24, Location 9). The core of the axis is largely covered by flat lying ultramafic and mafic rocks.

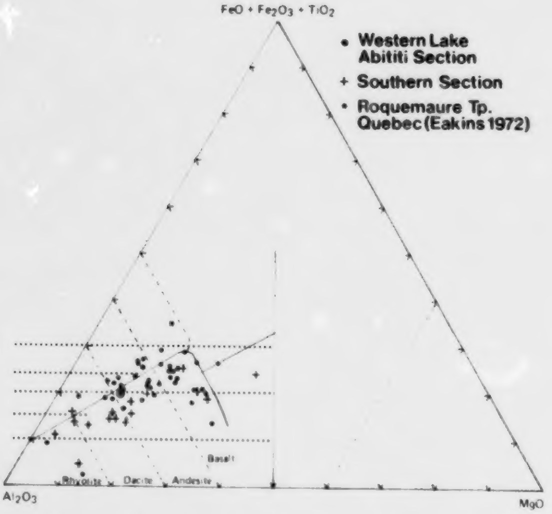


Figure 27. Chemical analyses of the Hunter Mine Group projected on the Jensen Cation Diagram (after Eakins 1972). The solid circle with concentric circle represents 2 samples from the western Lake Abitibi section which plot in the same position.

Rhyolite tuff is exposed along the southern edge of the ultramafic rocks, and in places, where deeply eroded, within the area of ultramafic rock, to suggest that the core of the syncline is mainly calc-alkalic dacite and rhyolite. Additional outcrops of calc-alkalic basalt are exposed in a few places near the southern edge of the fault block on the southern limb of the syncline.

Neither iron formation nor greywacke was observed with the calc-alkalic volcanic rocks in the above-mentioned area. However, minor graphite and cherty tuff have been reported from diamond drilling in Frecheville Township.

Bedrock is not exposed in the southwestern part of Lamplugh Township. Low aeromagnetic expressions in the area suggest that the area is largely underlain by calc-alkalic dacite and rhyolite, and that these rock extend westward into Rand Township (*see* Figure 21).

In southern Rand Township, a circular area 4.0 km in diameter, contains numerous poorly layered, calc-alkalic andesite, dacite, and rhyolite tuff-breccias cut by many dikes and sills of porphyritic andesite dacite and rhyolite (Figure 24, Location 10). The intrusions are 3 to 10 m across, and show complex intrusive relationships to one another. These bodies, as well as many of the breccia fragments, contain milky white plagioclase phenocrysts 2 to 5 mm long and quartz phenocrysts 0.5 to 1.5 mm in size set in a light grey to olive green matrix. The porphyries are indistinguishable from those found in the rhyolite complex described in the western Lake Abitibi section. This suggests that the circular area represents an explosive vent area. A few outcrops on the southern edge of the circular area show that the rhyolite complex grades southward into steeply dipping, finer grained schistose bedded tuffs.

There are no rock exposures north of the circular area shown on Figure 24. Farther to the west, in Indian Reserve 70 (Figure 24, Location 11), limited exposures indicate that the rocks are mainly andesite, dacite, and rhyolite tuff-breccias. A few pillowed and massive calc-alkalic basalt and andesite flows are interlayered with the breccias near the northern edge of the fault block.

The calc-alkalic rocks are thought to extend west into McCool, Warden, and Munro Townships where extensive areas are shown to be underlain by felsic volcanic rocks (Satterly 1952a, 1952b, 1953). Many of the mafic volcanic rocks shown by Satterly (1953) may be calc-alkalic basalts and andesites of the Hunter Mine Group as well. The work of Arndt (1975) shows that many of the volcanic rocks shown by Satterly (1953), are actually komatiitic and tholeiitic lavas. Additional mapping is required before the mafic volcanic rock types can be delineated throughout this area.

No iron formation has been observed or reported with the felsic tuffs, either in Rand Township, or farther west. Iron formation occurs with cherty tuffs and tuff-breccias in the northern part of Garrison Township (Figure 24, Location 12). Here, rocks of the Hunter Mine Group are overlain to the north by ultramafic rocks that dip north. This information comes from diamond-drill holes put down in search of asbestos deposits. Little is known about the extent of the iron formation or the felsic tuffs. However, the iron formation was intersected in 2 widely spaced holes, which suggests that this part of the group probably represents a depositional facies nearly equivalent to that found in Palmarolle Township.

Major element analyses were done on 40 rock specimens from the southwestern section of the Hunter Mine Group. Some of the specimens were from diamond-drill core which intersected the iron formation in the northern part of Garrison Township. Projection of these analyses on Figures 25, 26, and 27 cluster with projections of chemical analyses of rock specimens of the group from elsewhere. Figure 26 shows that the Hunter Mine Group is a calc-alkalic volcanic suite.

Sedimentary Rocks

Satterly (1949a, 1949b, 1952b, 1954) mapped narrow, but persistent zones of sedimentary rock north of the southern branch of the Destor - Porcupine Fault Zone (Figure 24, Location 14). Satterly (1954) described the problems associated with determining the stratigraphic position of these rocks. He suggested that the rocks in Harker, Garrison, and Michaud Townships were extensions of the sedimentary rocks farther west in Guibord, Munro, and Beatty Townships (Porcupine Group; Figure 24, Location 13). In Holloway Township, Satterly (1954) also found conglomerate in which pebbles of jasper iron formation and pink feldspar porphyry are abundant. This suggests that some sedimentary rocks might be Timiskaming-type rocks. Many zones of similar conglomerate, some associated with trachytic lavas and pink intrusive feldspar porphyries, were found elsewhere in diamond drilling along the southern branch of the Destor - Porcupine Fault Zone. In this study, these rocks are considered to be part of the Destor - Porcupine Complex (*see* Figure 19) and distinct from the exposed sedimentary rocks in Garrison and Harker Townships.

If the exposed sedimentary rocks in Garrison and Harker Townships are extensions of the Porcupine Group, as suggested by Satterly (1954), they would have to be considered to be as old or older than the ultramafic and mafic volcanic rocks as was suggested by Satterly and Armstrong (1949), Satterly (1952b), and Prest (1953) and be equivalent in age to the Hunter Mine Group. Therefore, such rocks would be a sedimentary facies of the Hunter Mine Group which occurs farther north.

Satterly (1952b, p.7) described the sedimentary rocks of Harker and Garrison Townships as being mainly greywackes interlayered with iron formation. The greywackes were fine-grained, grey, green or black rocks which contained minute grains of quartz visible only under a hand lens. He further described the greywackes as having angular quartz and feldspar (oligoclase) grains and rock fragments of quartzite or acid lava which were set in a groundmass of quartz, feldspar, chlorite, biotite, and white mica. This strongly favours a volcanic terrain having abundant dacite and rhyolite as a source of material for the greywackes, rather than a volcanic terrain having komatiitic and tholeiitic lavas.

The greywackes in Garrison Township occur as beds 10 to 200 cm thick. In a few cases, graded bedding can be observed. Here, the grain size decreases from sand to clay size toward the top of the beds where the beds become laminated. The tops of many beds are channeled by the overlying bed and indicate that the greywackes were deposited by turbidity currents (Turner and Walker 1973).

Satterly (1949b, p.7; 1952b, p.10) described the iron formation as being finely laminated and forming zones from a few feet to 100 feet thick within the sedimentary rocks. The iron formation referred to by Satterly (1949, 1951) occurs in beds 1 to 5 cm thick. The beds are of sand- to clay-size material which grades rapidly into iron oxide, mainly magnetite toward their tops so that the rocks appear as banded argillite and magnetite. Chert is rare or absent.

Goodwin and Shklanka (1967) implied that iron formation becomes interbedded with clastic material deeper in the basin adjoining a felsic volcanic terrain. Dimroth *et al.* (1973) concluded that argillaceous iron formations represent marine distal turbidites; he implied that these rocks were formed in a deeper basinal environment than was the cherty iron formation which he considered to have formed mainly in epi- and peri-cratonic basins. This would suggest that the sedimentary rocks of Harker and Garrison Townships are representative of a marine facies of the Hunter Mine Group.

Geological Interpretation

The Hunter Mine Group in the Lightning River area appears to have been deposited on the southern side of a volcanic pile, the early history of which is not known. The distribution of the rock suggests a transition from a volcanic environment to a marine environment (Figure 28). Towards the Lake Abitibi Batholith, the rocks are mainly calc-alkalic andesite dacite

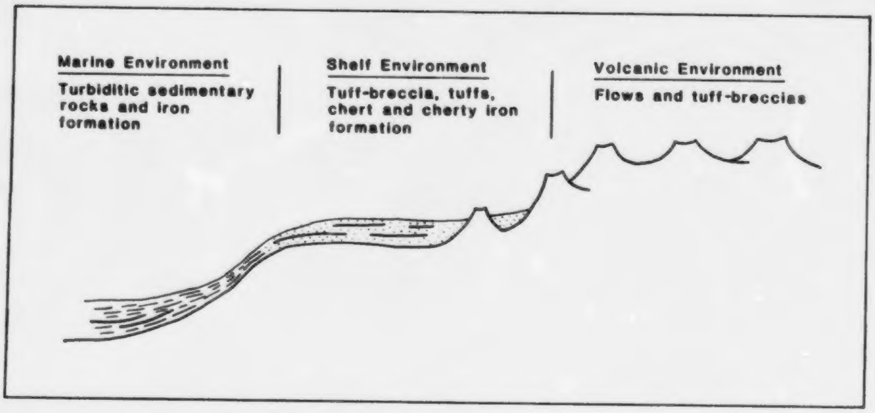

Marine Environment
Turbiditic sedimentary rocks and iron formation
Shelf Environment
Tuff-breccia, tuffs, chert and cherty iron formation
Volcanic Environment
Flows and tuff-breccias

Figure 28. Diagram of the transition from a volcanic environment to a marine environment.

and rhyolite breccias interlayered with some flows of calc-alkalic basalt and andesite. In these areas, greywacke and iron formation are absent. Elliptical to circular areas containing concentrations of intrusive dacite and rhyolite disrupt the volcanic breccias. These areas were probably explosive vents, but, instead of being main vents with thick piles of volcanic lava around them, they were probably satellite vents on the perimeter of a larger central vent area located in the area now underlain by the Lake Abitibi Batholith.

Away from the Lake Abitibi Batholith, the volcanic pile sloped southwestward and eastward where the volcanic rocks were deposited in deeper water. In these areas, the northern parts of Garrison and Palmarolle Townships, the felsic volcanic breccias and tuffs become interlayered with cherty iron formation and greywacke. This would suggest that these rocks were deposited in epi- and peri-cratonic - like environments (Dimroth 1975a, 1975b).

Turbidites composed of greywackes and iron formation are exposed farther to the south in Garrison and Harker Townships. This suggests that the volcanic pile graded rapidly into a marine basin along the present location of the middle branch of the Destor - Porcupine Fault Zone (*see* Figure 24). The thickness of these turbidites is unknown because of faulting and poor exposure, but if they are extensions of the sedimentary rocks farther westward (Porcupine Group), as suggested by Satterly (1954), a sequence of turbidites a few thousand metres thick (Prest 1957) may have been deposited south of the volcanic pile represented by the Hunter Mine Group.

The preservation of the Hunter Mine Group in a large fault block between the northern and middle branches of the Destor - Porcupine Fault Zone suggests that the volcanic pile became segmented near its margin. A fault probably developed and the 2 sections were pulled apart so that mafic and ultramafic lavas could fill the area between the northern branch fault and the Lake Abitibi Batholith. Also, the main volcanic pile to the north and the rhyolite complex in Rand Township may have had a valley between them which became flooded by basaltic lavas and was downfaulted.

PORCUPINE GROUP

The sedimentary rocks (Figure 24, Location 13) can be traced west into the Timmins area where they have been described and named the Porcupine Group (Lorsong 1975, and Pyke 1980). Pyke (1980) considered these sedimentary rocks to be equivalent to the Upper Formation of the Deloro Group and the Tisdale Group (Figure 29). Lorsong (1975) and Pyke (1982) described the rocks as being mainly turbiditic greywackes and argillites. These rocks appear to lack iron formation. In places, the turbidites are overlain by fluviatile sedimentary rocks.

In the Timmins area, the Porcupine Group is largely confined to the northern side of the Destor - Porcupine Fault Zone and does not extend far south of Timmins (*see* Figure 5). Three possible reasons can be suggested for the lack of sedimentary rocks south of Timmins: firstly, the Porcupine Group was not deposited in the area south of Timmins; secondly, the group was deposited, but was removed by later erosion; or thirdly, the group was deposited and covered by younger mafic volcanic rocks belonging to the Tisdale Group which Pyke (1982) correlated with the Stoughton - Roquemaure, Kinojevis, and Blake River Groups in the Kirkland Lake area. The latter suggestion is favoured by the author, but this would also imply that the Porcupine Group is only equivalent to the Upper Formation of the Deloro Group.

UPPER FORMATION OF THE DELORO GROUP

The Upper Formation of the Deloro Group occurs mainly south of Timmins (Pyke 1982). Pyke (1978a, 1978b, 1980) described the Upper Formation as being mainly composed of calc-alkalic dacite and rhyolite breccias. It has been dated at 2725 ± 2 Ma (Nunes and Pyke 1980), and overlies calc-alkalic basalts and andesites in the Shaw Dome just south of Timmins (*see* Figure 5). Farther to the south, the Upper Formation also overlies tholeiitic basalts and minor komatiites which occur close to the Kenogamissi Batholith. Pyke (1982) referred to these lower rocks as the Lower and Middle Formations of the Deloro Group.

Pyke (1978a, 1978b, 1982) described the breccias of the Upper Formation as being poorly layered and containing quartz-feldspar porphyry fragments. Toward the top of the Upper Formation, iron formation

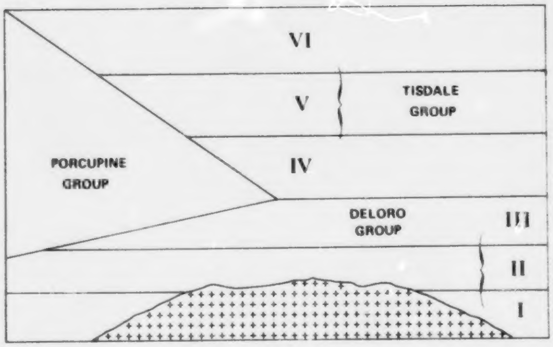


Figure 29. Diagrammatic illustration of time equivalence of sedimentation (Porcupine Group) and volcanism (Deloro and Tisdale Groups) in the Timmins area (after Pyke 1982).

consists of alternating layers of magnetite and chert. In places, the iron formation grades into a sulphide facies in which graphite and argillite are abundant (Pyke 1982).

Pyke (1982) used the iron formation as a marker horizon for interpreting the stratigraphy south of Timmins. This suggests that the Upper Formation shows only 1 facies of deposition in this area; and that even though it may have been tilted, it had not been rotated in 2 directions to show other facies of deposition. The exposed facies of the Upper Formation suggests that it was deposited near the edge of a basin (Dimroth 1975a, 1975b) which was present to the east. This facies of deposition was probably similar to that of the Hunter Mine Group found in northern Garrison and Palmarolle Townships. A succession of turbidites equivalent to the Porcupine Group was probably deposited in the basin to the east. The rocks on the southern side of the Destor - Porcupine Fault must have been displaced downward, so that the marine facies of the Upper Formation (Porcupine Group) is not exposed.

Since older more mafic volcanic rocks are present below the Upper Formation, this would suggest that it represents the top of a volcanic cycle as suggested by Pyke (1980). The Upper Formation as well as the lower rocks appear mainly to dip steeply eastward. This suggests that either these rocks were tilted inward toward the basin as the basin was being filled with mafic volcanic rocks and sagged, or that the rocks were tilted outward from the Kenogamissi Batholith as the batholith domed to the west.

SKEAD GROUP

The Skead Group, referred to as the Skead Pyroclastics by Ridler (1969), has been described by Hewitt (1951) and Ridler (1969) as being dominantly heterogeneous felsic pyroclastic rocks. According to Ridler (1969), the Skead Group is up to 23 000 feet thick east of the Round Lake Batholith (*see* Figure 5) and thins to the northwest. Hewitt (1951) and Ridler (1969) suggested that much of the volcanic material was derived from a volcanic centre located near to the top of the Skead Group west of the Round Lake Batholith. Both authors also suggested that trachyte lava flows occurred near the top of the Skead Group. This has not been confirmed by the author, nor do the chemical analyses of Ridler (1969) confirm this observation. However, the senior author has observed a conglomerate at the top of the Skead Group which contains trachyte pebbles with more than 4% K_2O. Most of the chemical analyses of rhyolite from the top of the Skead Group have more than 6.5% N_2O and less than 1.5% K_2O (Jensen 1978c).

Hewitt (1951) noted the presence of quartz phenocrysts of 3 to 5 mm in size in the felsic rocks. This size of quartz phenocrysts is diagnostic of the Skead Group throughout the area. To date, the author has not observed this size of quartz phenocrysts in rhyolite elsewhere in the Abitibi Greenstone Belt.

Iron formation has not been reported in the Skead Group, except to the northwest, near the Otto Syenite Stock (Lovell 1972). Here, the Skead Group becomes thin and is composed mainly of cherty tuffs (*see* Figure 5). This indicates that in this area, the Skead Group was probably deposited in deeper water and that volcanic activity was limited. The thickening and coarsening of tuff and tuff-breccia eastward around the batholith and of the underlying mafic rocks (*see* Figure 5) suggests that the thickest part of the Skead Group has been rotated eastward through a 90° arc from an upright position to a sideways position so that a cross section of a volcanic pile is presently exposed. The pivotal line would have been a north-south line through the Round Lake Batholith and the Otto Syenite Stock. However, no sedimentary rocks that may have formed in deeper water appear to have been rotated up toward the west. Sedimentary rocks may have been deposited in a basin to the north, and may not have been part of the rotation. This will be discussed further in the report.

A thick sequence of tholeiitic basalts called the Catherine Group (Catherine basalts by Ridler 1969) as well as a group of komatiitic lavas (Wabewawa Group) underlie the Skead Group. These mafic to ultramafic lavas overlie an older group of calc-alkalic rocks called the Pacaud Tuffs (Ridler 1969).

PACAUD TUFFS

The Pacaud Tuffs have been described by Ridler (1969, 1975) as encircling the Round Lake Batholith and as being overlain by mafic lava flows. Ridler (1975) also described a thin persistent unit of iron formation in the tuffs, a feature which had been noted earlier by Moorhouse (1944) and Lawton (1959) on the southern and northern sides of the batholith, respectively.

Both cherty oxide and argillaceous sulphide facies iron formation are present. These facies occur interlayered with calc-alkalic dacite and andesite tuffs and tuff-breccias. Greywacke and argillite are also present. As noted by Ridler (1969, 1975), these rocks are strongly foliated, and except for the iron formation, are difficult to distinguish and to delineate for any distance.

The persistence of the iron formation suggests that the Pacaud Tuffs show only 1 facies of their deposition. Also, this facies of deposition occurred in a relatively deep, quiet-water environment at the edge of a volcanic pile. By analogy with the Hunter Mine

Group and the Lake Abitibi Batholith, the centre of the volcanic pile would have been located in the area presently occupied by the Round Lake Batholith.

A thick apron of sedimentary rock was probably deposited in the basinal area which surrounded the volcanic pile represented by the Pacaud Tuffs. A part of this basin would have been to the north, and would have joined the basin south of the volcanic pile represented by the Hunter Mine Group and the basin situated east of the Upper Formation, Deloro Group.

SIGNIFICANCE OF THE CALC-ALKALIC VOLCANIC ROCKS OF THE LOWER SUPERGROUPS

The Hunter Mine and Skead Groups, the Upper Formation of the Deloro Group, and the Pacaud Tuffs represent late-stage volcanism of separately developed volcanic piles situated around the margins of a basin (*see* Figure 23). The Upper Formation, Deloro Group (2725 ± 3 Ma) occurs between the Destor - Porcupine Fault Zone and the Kirkland Lake - Larder Lake Fault Zone at the western end of the synclinorium (*see* Figure 5). This suggests either that the 2 fault zones formed separately, or, that they formed one suture which widened to allow volcanic piles to form between them, as suggested by Pyke (1980). This must have happened prior to development of the volcanic pile represented by the Deloro Group. The Hunter Mine Group (2709 ± 2 Ma) and the Skead Group (2709 ± 2 Ma) formed later, suggesting that the faults were not developed prior to the deposition of the Deloro Group.

Rocks of the Skead Group are characterized by large quartz phenocrysts. The Hunter Mine Group rocks have large feldspar phenocrysts and small quartz phenocrysts, implying that the 2 groups of rocks were not 1 group that was later split into 2 parts. Instead, the Hunter Mine Group grades southward into sedimentary rocks which are probably correlative with the Porcupine Group. The rocks of the Porcupine Group extend west and become more abundant in the Timmins area. This also suggests that the Hunter Mine Group represented one of a chain of volcanic piles developed north of the Destor Porcupine Fault Zone.

The Skead Group represents the top of a volcanic pile turned on its side and appears to have had a history quite different from that of the Hunter Mine Group or the Pacaud Tuffs. The rocks of the Pacaud Tuffs show little change around the Round Lake Batholith. This suggests that they represent only one depositional facies of felsic volcanism. Even though the Pacaud Tuffs may have been tilted, they have not been rotated, so that they show more than one depositional facies, as does the Skead Group.

Goodwin and Shklanka (1967), concluded that iron formation signified the transition from a volcanic environment to a sedimentary environment. The presence of iron formation in the Hunter Mine Group, Upper Formation of the Deloro Group, and Pacaud Tuffs thus indicates that a thick wedge of sediments may have been formed around the periphery of a basin in the area now occupied by the synclinorium (*see* Figures 23 and 5). Except for the Porcupine Group, these sedimentary rocks are not exposed.

The calc-alkalic rocks of the older volcanic piles are mainly rhyolitic and dacitic tuff-breccia and tuff which have quartz phenocrysts. This would imply that the magmas for these rocks were generated at depths of about 100 km (Green and Ringwood 1968) by the partial melting of eclogite or garnet-bearing amphibolite of intermediate composition (Ewing 1979).

KOMATIITIC VOLCANIC ROCK SEQUENCES

In this report, the komatiitic sequences are groups of volcanic rocks in which tholeiitic lavas, and in places, calc-alkalic felsic tuff and tuff-breccia and turbiditic sedimentary rocks are interlayered with the komatiitic volcanic rocks. These groups of rock form the basal parts of supergroups which contain additional groups of tholeiitic, calc-alkalic, and alkalic volcanic rocks in which komatiitic lavas are absent (*see* Figure 5). The komatiitic lavas were among the first components of each cycle of volcanism because they are at the base of supergroups.

Sun and Nesbitt (1978a) and Nesbitt *et al.* (1979) emphasized that the tectonic environment in which komatiites are formed is still unknown. In the Kirkland Lake area, the komatiitic sequences overlie the calc-alkalic volcanic rocks which are mainly late members of volcanic piles, and were deposited near or at the transition from volcanic to sedimentary sequences. This indicates that the komatiitic sequences were deposited at the edges of earlier formed volcanic piles and filled low areas among them.

The main komatiitic sequence of the Kirkland Lake area occurs at the base of the Upper Supergroup (*see* Figure 5), and, on the northern limb of the synclinorium forms the Stoughton - Roquemaure Group (Jensen 1976a, 1978b). In the Kirkland Lake area, and near Timmins, it is the Bowman Subgroup (Goodwin 1977) and the Lower Formation of the Tisdale Group (Pyke 1980). On the southern limb of the synclinorium, south of Kirkland Lake, the komatiitic sequence, called the Larder Lake Group (Jensen 1978c), includes rocks previously named the McVittie Basalts, the Highway 11 Basalts, and the Timiskaming Group (Ridler 1969, 1970).

Komatiitic sequences present near the base of the lower supergroups south of Kirkland Lake are the Wabewawa Group (Jensen 1978c) and south of Timmins, the Lower Formation of the Deloro Group (Pyke 1980; *see* Figure 5).

The sequence of rocks in the Stoughton - Roquemaure Group illustrates the development of the komatiitic volcanic rock sequences of the whole Kirkland Lake area. This sequence is similar to those in Munro Township, Dundonald Township, as well as elsewhere in Ontario and provides a basis for understanding the regional tectonic environment of the komatiitic sequences.

STOUGHTON - ROQUEMAURE GROUP

Distribution

The Stoughton - Roquemaure Group occupies much of the area north of the Destor - Porcupine Fault Zone (Figure 30). The only place where both a definable base and top were seen, is in the type section of the group in the townships of Roquemaure and Stoughton (Jensen 1976a, 1978b) (Figure 30, Locations 1 and 2). The Stoughton - Roquemaure Group disconformably overlies the felsic volcanic rocks of the Hunter Mine Group and is overlain by thin horizons of felsic volcanic rock and iron formation. The iron formation and felsic volcanic rocks are overlain by tholeiitic lavas belonging to the Kinojevis Group in which komatiitic lavas are absent. The Stoughton - Roquemaure Group type section is well exposed along the southern shore and adjacent islands of Lake Abitibi. From here, the group extends westward across the length of Upper Lake Abitibi into Milligan and Warden Townships, where rock exposures become poor and are largely unmapped. The part of the Stoughton - Roquemaure Group at the western end of Lake Abitibi is referred to as the "Western Lake Abitibi Section".

Basal sections of the Stoughton - Roquemaure Group are located in the southern parts of Lamplugh and Frecheville Townships (Ghost Range section), along the northern boundary of Garrison Township (Garrison section), and along the northern side of the southern branch of the Destor - Porcupine Fault Zone (Destor - Porcupine section). Together, these basal sections, represent a zone of komatiitic lavas which extend west through Guibord and McCool Townships into Munro Township where these rocks have been described in detail by Arndt (1975) and Arndt *et al.* (1977).

Type Section (Roquemaure Township)

The type section of the Stoughton - Roquemaure Group is 11 km thick, and forms the lower part of a monoclinal volcanic succession, that is more than 15 km thick, and faces southwest with a dip of 80 to 90 degrees (Figure 30). The lower 5 km of the group is located in Roquemaure Township, Quebec, where it has been described by Eakins (1972).

In the lower 1500 m of the Stoughton - Roquemaure Group, Eakins (1972) found elongated, coarse-grained bodies of gabbro, peridotite, and dunite interlayered with lesser amounts of massive and pillowed lava above which the rocks were mainly fine-grained massive and pillowed lavas. Eakins (1972) also noted that there were a few finely bedded felsic tuff horizons interlayered with the mafic and ultramafic rocks. The peridotites and gabbros near the base of the Stoughton - Roquemaure Group cut at low angles, both upper Hunter Mine Group rocks and the lowermost mafic and ultramafic lavas of the Stoughton - Roquemaure Group. (Eakins 1972). Above the base, the gabbros according to Eakins seemed to be largely conformable with, and in places, seemed to grade into the enclosing lavas. Eakins implied that many of these gabbros could have been massive flows.

At the base of the group, Eakins (1972) described the peridotites as intrusions distinct from the gabbro intrusions. The largest body of peridotite was 160 m wide, and greater than 2 km long. It consists mainly of serpentine pseudomorphs after rounded olivine grains with some magnetite and pyrite (Table 7, Num-

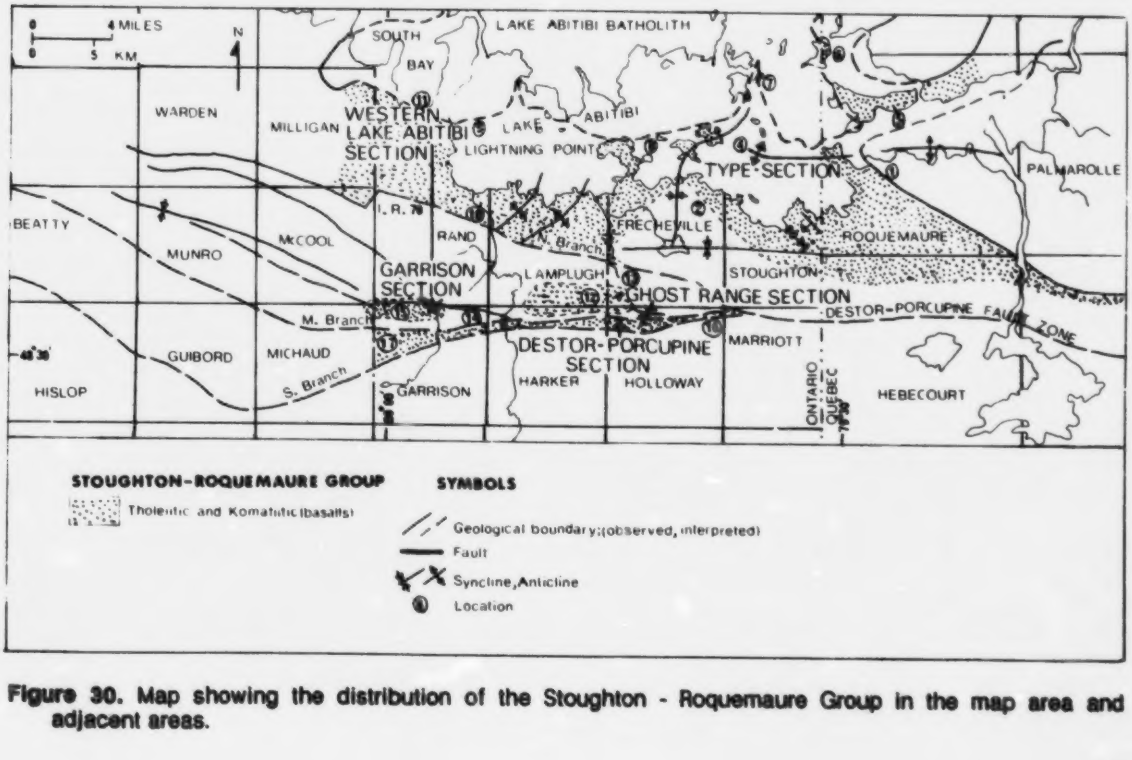


Figure 30. Map showing the distribution of the Stoughton - Roquemaure Group in the map area and adjacent areas.

Table 7

TABLE 7: SEMI-QUANTITATIVE CHEMICAL ANALYSES OF INTRUSIVE ROCKS FROM THE STOUGHTON - ROQUEMAURE GROUP IN ROQUEMAURE TOWNSHIP (AFTER EAKINS 1972).

SAMPLE NUMBER	18	19	20	21	22	23	24	25
SiO_2	37.80%	37.72%	44.12%	47.30%	45.29%	47.52%	47.13%	47.82%
TiO_2	0.08	0.13	1.49	2.40	2.62	1.44	1.23	1.44
Al_2O_3	2.75	3.94	12.34	11.28	11.20	13.87	13.35	13.48
CaO	0.20	3.54	9.64	8.43	9.55	9.40	8.95	11.13
MgO	36.94	32.46	4.06	3.99	3.94	5.57	8.03	6.42
FeO*	6.36*	9.31*	12.73	18.95*	21.37*	15.02*	14.19*	13.25*
Na_2O	0.08	0.07	2.70	3.06	1.86	2.66	1.78	2.21
K_2O	0.02	0.06	0.06	0.19	0.26	0.05	0.07	0.14
V	0.010	-	0.033	0.031	0.045	0.031	0.030	0.026
Cr	0.026	-	0.007	0.015	0.005	0.008	0.016	0.007
Fe (total)	4.93	7.22	10.52	14.69	16.57	11.64	11.00	10.27
Ni	2640 ppm	2160 ppm	52 ppm	22 ppm	25 ppm	64 ppm	79 ppm	82 ppm
Cu	26 ppm	25 ppm	94 ppm	66 ppm	112 ppm	105 ppm	139 ppm	120 ppm
Zn	47 ppm	46 ppm	115 ppm	120 ppm	105 ppm	115 ppm	73 ppm	95 ppm
Pb	13 ppm	12 ppm	15 ppm	16 ppm	13 ppm	32 ppm	13 ppm	18 ppm

Notes:

*Original determination was for total Fe. For purposes of calculation and comparison, total Fe has been entirely converted to FeO.

18. Serpentinite derived from a dunite; southern end of lot 48, range I, in main asbestos zone (Analysis by the Laboratory Branch, Quebec Department of Natural Resources Q.D.N.R.).
19. Serpentinized dunite: southern end of lot 48, range I, in the main asbestos zone (Analysis by the Laboratory Branch, Q.D.N.R.).
20. Altered gabbro; northern end of lot 14, range II, about 700 feet south of the range II-III road (Analysis by the Laboratory Branch, Q.D.N.R.).
21. Gabbro from an exposure at the southern end of lot 48, range III (Analysis by the Laboratories Branch, Q.D.N.R.).
22. Gabbro from the southern end of lot 52, range III (Analysis by the Laboratory Branch, Q.D.N.R.).
23. Diorite from an exposure 650 feet south of range II-III road in lot 19, range II.
24. Early diorite, lot 23, range IV.
25. Amphibolitized diorite from shoreline exposure, southern shore of Nepawa Island, lot 30, range VIII.

bers 18 and 19). Eakins (1972) observed some pillowed lavas of undetermined composition overlying the peridotite, which suggests that this peridotite represents a uniform intrusion.

The 'Roquemaure Gabbro', near the base of the Stoughton - Roquemaure Group, is an elongate body several km long and 100 to 200 m wide, that Eakins described as being a medium-grained, dark green gabbro with a subophitic, hypidiomorphic granular texture composed of uralitized pyroxene, saussuritized plagioclase, quartz, and magnetite. According to Eakins (1972), chemical analyses showed the Roquemaure gabbro to be about 20% FeO_T and 4% weight (Table 7, Numbers 21 and 22). In many places, the gabbro is pegmatoidal and contains xenoliths of rhyolite and basaltic lava. The pegmatoid phases are not unlike those described in the gabbroic rocks of the Labrador Trough by Dimroth *et al.* 1970. Eakins (1972) showed that the Roquemaure gabbro intruded the uppermost rhyolitic rocks of the Hunter Mine Group and the basal lavas of the Stoughton -Roquemaure Group.

Eakins (1972) segregated the other gabbros into "red chert gabbro" and "gabbro and diorite". He described the red chert gabbro (Table 6, Number 20) as a group of gabbroic intrusions containing red, red and white, and black chert fragments, blocks, and septs up to 7 m long. These chert fragments were considered to be derived from the underlying felsic volcanic rocks of the Hunter Mine Group. In places, Eakins (1972) showed narrow bodies of rhyolite, up to 2 km long, within his "red-chert gabbro". The gabbro is fine to medium grained, olive greenish grey to dark green on fresh surfaces, and is composed of altered pyroxene, olivine, and plagioclase replaced by carbonate, serpentine, chlorite, magnetite, and in places, by quartz. Porphyritic varieties have pyroxene phenocrysts up to 2 mm long in a diabasic groundmass composed of clinopyroxene, plagioclase (An_{90} to An_{60}), and magnetite partly altered to epidote and chlorite.

Eakins (1972) indicated that the remaining gabbros and diorites are similar to the "red chert gabbro" except that they did not contain the red chert xenoliths. A reexamination of some of these rocks by the author, however, suggested that they are similar to thick lava flows found elsewhere in the Stoughton - Roquemaure Group and in the Kinojevis Group.

TABLE 8: SEMI-QUANTITATIVE CHEMICAL ANALYSES OF MAFIC TO ULTRAMAFIC LAVA FLOWS FROM THE STOUGHTON - ROQUEMAURE GROUP IN ROQUEMAURE TOWNSHIP (AFTER EAKINS 1972).

SAMPLE NUMBER	4	5	6	7	8
SiO_2	45.39%	41.51%	47.61%	48.48%	47.61%
TiO_2	0.72	0.50	0.68	0.67	1.24
Al_2O_3	7.53	10.70	8.41	14.76	13.80
CaO	9.12	9.21	10.28	11.57	11.25
MgO	18.94	16.84	14.59	7.46	6.13
FeO*	11.97	11.63	12.86	11.91	14.36
Na_2O	0.45	0.79	1.48	1.88	2.10
K_2O	0.11	0.03	0.07	0.04	0.44
V	0.017	0.015	0.016	0.016	0.030
Cr	0.022	0.27	0.23	0.057	0.006
Fe (total)	9.28	9.02	9.97	9.23	11.13
Ni	880 ppm	790 ppm	670 ppm	151 ppm	97 ppm
Cu	40 ppm	189 ppm	153 ppm	155 ppm	135 ppm
Zn	87 ppm	74 ppm	83 ppm	85 ppm	108 ppm
Pb	17 ppm	18 ppm	13 ppm	22 ppm	18 ppm

Notes:

* Original determination was for total Fe. For purposes of calculation and comparison, total Fe has been entirely converted to FeO.

4. Amphilobitized pillowed picrite basalt; sample taken from single pillow centre; shoreline exposure, northern shore of Nepawa Bay, lot 21, range IX, Nepawa Island.

5. Pillowed picrite basalt; whole rock sample taken from small road cut on western side of lot 20-21 road in the northern part of range I.

6. Pillowed picrite basalt; sample from single pillow from a large ridge exposure at the northern end of lot 19, range I.

7. Pillowed basalt; sample taken from centre of single pillow from clear shoreline exposure, lot 6, range IV.

8. Finely amphibolitized basalt with stretched pillows from road cut, lot 15, range VIII, Nepawa Island.

All analyses by the Laboratory Branch, Quebec Department of Natural Resources.

Except for the Roquemaure gabbro, the remaining gabbros which Eakins described as being sills, could be thick tabular lava flows. Elsewhere, such lavas have interflow cherty tuff, chert, and iron formation at their base, and have ripped up and incorporated parts of these interflow rocks as xenoliths. It is probable that the "red chert gabbros" and other gabbros of Eakins (1972) are equivalent to lava flows found elsewhere near the base of the Stoughton - Roquemaure Group.

In Roquemaure Township, Eakins did not report gabbros with cumulate layering of olivine, pyroxene or plagioclase, as described by MacRae (1969) or Arndt (1975). The high iron and titanium contents of the gabbros (Table 7) suggest that these rocks were derived from highly differentiated magmas caused by the crystal fractionation of pyroxene and plagioclase at shallow depths (Green and Ringwood 1967a).

The mafic and ultramafic lavas are composed of well-pillowed flows with scant associated breccia, and massive flows devoid of structure except for very narrow flow-contact zones a few inches thick (Eakins 1972). Pillowed layers, about 5 m thick, are interlayered with the massive flows and some pass laterally into massive flows. Eakins (1972) described some lavas as fine-grained, grey to green rocks with a mafic to ultramafic mineral and chemical composition. Others, he described as fine-grained, dark greyish green to greenish black basalts with serpentine pseudomorphs after equant, lath-shaped, and doubly terminated olivine phenocrysts. Eakins (1972) did not distinguish the ultramafic lavas from the mafic lavas on his map; therefore, it is difficult to estimate their proportions, but from his description many appeared to be peridotitic and basaltic komatiites. His chemical analyses showed 3 komatiites; a magnesium-rich tholeiitic basalt, and an iron-rich tholeiitic basalt (Table 8). This would suggest that pillowed tholeiitic basalts occurred with the komatiites in the lower part of the Stoughton - Roquemaure Group.

If many of the gabbroic rocks are thick lava flows, then a high proportion of the volcanic rocks

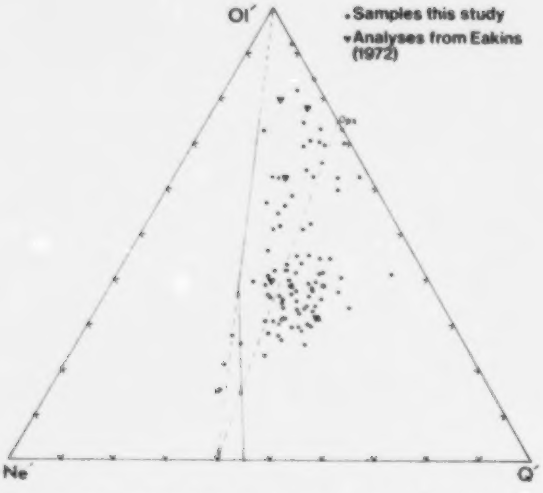


Figure 31. Ol'-Ne'-Q' projections of rock analyses from the upper 6 km of the Stoughton - Roquemaure type section. Inverted black triangles represent analyses from Eakins (1972); (*see* Table 7).

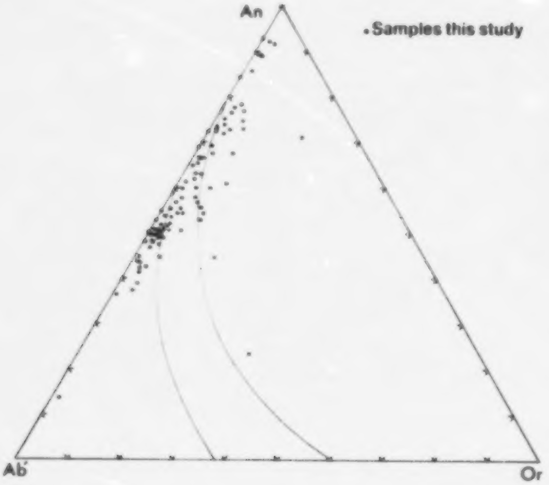


Figure 32. An-Ab'-Or projections of rock analyses from the Stoughton - Roquemaure Group type section.

near the base of the group must be iron-rich tholeiitic basalts with some magnesium-rich tholeiitic basalts associated with them. This would suggest that tholeiitic lavas erupted immediately before the bulk of the komatiitic lavas. If the present erosional level, however, is at a relatively high point on the paleoslope of the felsic volcanic pile, considerable amounts of both komatiitic and tholeiitic lava could have been extruded at lower levels before deposition of the rocks presently observed.

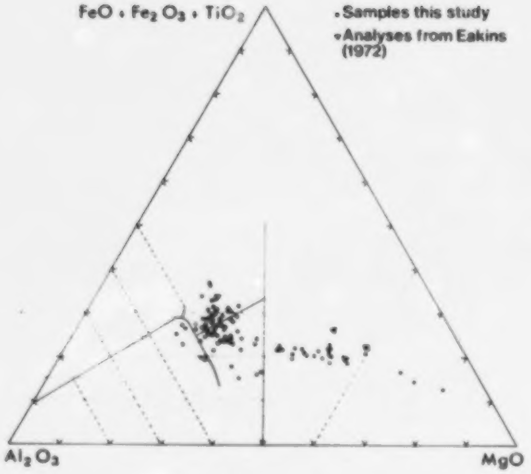


Figure 33. Projections of the chemical analyses of the volcanic rocks of the Stoughton - Roquemaure Group type section on the Jensen Cation Diagram. Inverted black triangles are from Eakins (1972); (*see* Table 7).

Type Section (Stoughton Township)

The upper 6 km of the Stoughton - Roquemaure Group in Stoughton and Frecheville Townships (*see* Figure 30) has been described by Jensen (1978b). Eighty-six chemical analyses plotted on Ol'-Ne'-Q', An-Ab'-Or, and Jensen Cation Diagram projections (Figures 31, 32, and 33) show that the upper 6 km of the Stoughton -Roquemaure Group is largely composed of subalkalic, orthoclase-poor lava flows that range in composition from peridotitic komatiite to iron-rich tholeiitic basalt. Interflow felsic tuffs, cherts, and iron formations were not chemically analyzed.

Fourteen volcanic units form the upper 6 km of the Stoughton - Roquemaure Group, each of which is composed of only 1 rock type and consists of 1 or more flows of massive and pillowed lavas (Jensen 1976a, 1978b). The units form a cyclical pattern of peridotitic komatiite at the base overlain consecutively by basaltic komatiite, magnesium-rich tholeiitic basalt, and iron-rich basalt. These cycles vary from 300 m to 1700 m in thickness, and can have oscillations from 1 rock type to another, or can be incomplete, lacking iron-rich tholeiitic basalt at their tops or peridotitic komatiite at their bases. The units of ultramafic komatiite and basaltic komatiite generally decrease in thickness and abundance, whereas magnesium-rich tholeiitic basalt increases upward in the higher cycles. Some cycles are separated by thin units of iron formation and chert less than 10 cm thick.

Polyhedral jointing occurs in the komatiitic lava flows, but well developed olivine and clinopyroxene spinifex textures as described by Pyke *et al.* (1973) and Arndt *et al.* (1977) in Munro Township were not observed here, except in thin sections taken from near the margins of pillows and flows.

The iron-rich tholeiitic basalt flow at the top of the Stoughton - Roquemaure Group is 350 m thick and is composed of a dark green, diabasic textured

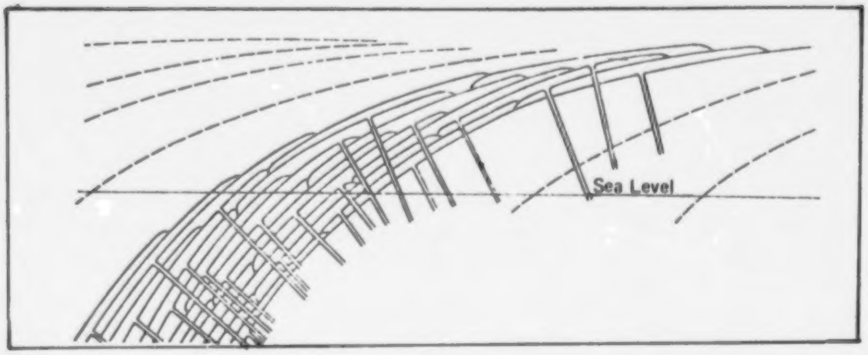


Figure 34. Progressive increase in dip of basalt lavas in Iceland towards central graben (redrawn from Walker 1965, Figure 2).

rock with grains 2 to 5 mm in size similar to the Roquemaure Gabbro described by Eakins (1972). The high aeromagnetic expression of this flow can be observed on Figure 21. In Figure 33, the 3 highest points in the iron-rich tholeiitic basalt field represent this lava flow.

Except for the upper iron-rich tholeiitic basalt flow and the gabbroic rock of Eakins (1972; Table 6, and Figure 37), the Stoughton - Roquemaure Group generally does not show a pronounced iron-enrichment trend. Instead, it shows mainly a trend of aluminum enrichment and magnesium depletion that extends from the peridotitic komatiite to the magnesium-rich tholeiitic field (see Figure 33). This trend was caused possibly by olivine fractionation of peridotitic magmas which resulted in the extrusion of basaltic komatiite, and magnesium-rich tholeiitic basalt because olivine fractionation would not be expected to cause an iron-enrichment (Donaldson 1975). Only after the magma had reached the composition of magnesium-rich tholeiitic basalt could plagioclase precipitate sufficiently to produce the limited iron enrichment seen in the rocks of the Stoughton - Roquemaure Group. The chemical trends produced by the rock analyses of Theo's Flow and Fred's Flow (Arndt 1975) are similar to that found in the rocks of the type section of the Stoughton - Roquemaure Group.

The alternative hypothesis to olivine fractionation of peridotitic magma would be that the lavas represented magmas from different stages of partial melting of a mantle diapir (Whitford and Arndt 1978), or that the magmas for each rock type were derived by partial melting of different magma sources (Nesbitt *et al.* 1979). Either of these 2 models should produce a discontinuous trend on Figure 33 and be recognizable in the stratigraphy. Additional analyses of the rocks in the lower part of the Stoughton - Roquemaure Group complemented by trace element studies, however, are required to test fully the 3 models suggested previously.

The cherts and cherty tuffs intercalated with the mafic and ultramafic lavas of the Stoughton - Roquemaure Group appear to have been derived from calc-alkalic volcanism elsewhere. They may represent waning phases of the volcanism that produced the Hunter Mine Group, or the erosional products of such rocks developed earlier. The presence of these felsic volcanic rocks with the mafic to ultramafic lavas may correspond to bimodal volcanism observed in other greenstone belts by Barker and Peterman (1974), Shaw (1976), Glikson and Lambert (1976), and Glikson (1976c).

The volcanic rocks of the Stoughton - Roquemaure Group in its type section all dip 80 to 90° to the southwest. As pointed out by Eakins (1972), the rocks in the type section are undeformed and except for their steep dip, show no evidence of folding. In addition, nothing indicates that the rocks were metamorphosed more at their base than at the top. Like basal rocks elsewhere (Jolly 1978), the basal rocks of this section were never deeply buried. If a 15 km vertical pile of lavas had been developed on top of the basal lavas prior to their being tilted nearly 90°, some evidence of metamorphism and compression should be present. On the other hand, there is little or no evidence to suggest that the lavas were not tilted as a single unit.

Walker (1965) and Bridgwater *et al.* (1974) have suggested that lava flows can become steeply tilted, during their deposition, toward a central graben, as shown in Figure 34. According to this model, the lavas can be dipping as steeply toward the top of the succession as they do at the bottom of the succession at sea level in the figure; and subvolcanic feeders for the lavas would only be observed toward the base of the lava succession.

Western Lake Abitibi Section

From its type section, the Stoughton - Roquemaure Group strikes northwest towards the Lake Abitibi Batholith. An anticline occurs to the north (Figure 30, Location 4), and the lower part of the group is deflected east toward Roquemaure Township (Figure 30, Location 5). In Roquemaure Township, the Stoughton - Roquemaure Group becomes highly amphibolitized and deformed by the Nepawa granitic body which is an extension of the Lake Abitibi Batholith (Eakins 1972). From here, the rocks extend around the granitic body and re-enter Ontario to form the bedrock underlying the north part of Kenosha Island (Figure 30, Location 6). The remaining upper part of the group becomes strongly altered and deformed into a tight north-trending syncline that pinches out into the granitic rocks to the north (*see* Figure 30, Location 7) (*see* Map P.2433, back pocket).

On the western limb of the syncline, the rocks strike west to southwest away from the margins of the

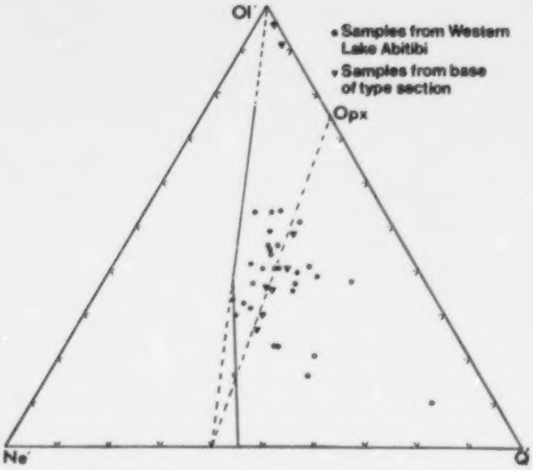


Figure 35. Ol′-Ne′-Q′ projections of analyses of basal coarse-grained rocks of Stoughton - Roquemaure Group (*see* Table 6, after Eakins (1972).

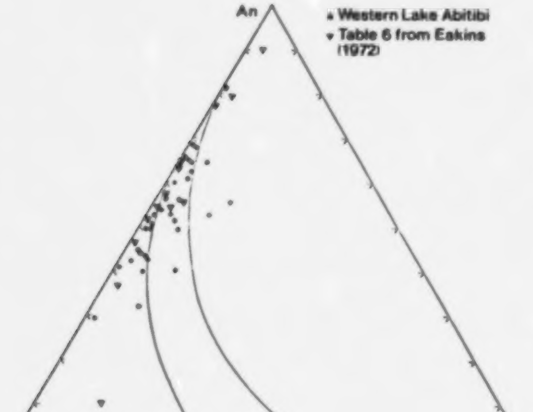


Figure 36. An-Ab′-Or projections of analyses of the coarse-grained basal rocks of the western Lake Abitibi and type sections of the Stoughton - Roquemaure Group (*see* Table 6 after Eakins 1972).

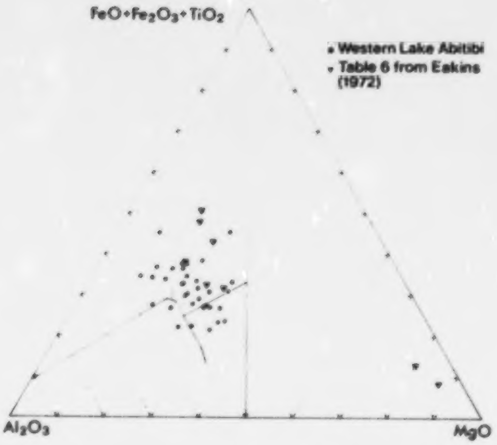


Figure 37. Jensen Cation Diagram projections of chemical analyses of basal coarse-grained rocks of the Stoughton - Roquemaure Group from the western part of Lake Abitibi and Roquemaure Township. Black circles are western Lake Abitibi and inverted triangles are from Eakins (1972).

Lake Abitibi Batholith, and become less altered and deformed as distance increases from the batholith. In this area, the base of the Stoughton - Roquemaure Group remains cut off by the batholith (*see* Figure 30, Location 8). West of Lightning Point, the group is exposed in the western part of Upper Lake Abitibi where the lower levels of the group are preserved. In the area, gabbroic rocks at the base of the group and rocks of the Hunter Mine Group are exposed (*see* Figure 30, Location 9). To the south, the upper parts of the Stoughton - Roquemaure Group become truncated by the northern branch of the Destor - Porcupine Fault as the upper parts of the group extend westward (*see* Figure 30, Location 10).

The Stoughton - Roquemaure Group appears to be about 7 km thick at the western end of Upper Lake Abitibi. The rocks are mainly south facing, with dips of 50 to 60% near the base of the section, and 80 to 90% in the middle and upper parts of the section.

The basal contact of the Stoughton - Roquemaure Group with the Hunter Mine Group is exposed in 1 outcrop on the southern shore of South Bay, Lower Lake Abitibi (*see* Figure 30, Location 11). Here, Hunter Mine Group rhyolite is overlain by a pillowed flow of magnesium-rich tholeiitic basalt of the Stoughton - Roquemaure Group which, to the west, also overlies coarse-grained, iron-rich tholeiitic basalt to indicate that the basal contact represents a disconformable contact (Map P.2433, back pocket).

At the western end of Upper Lake Abitibi (*see* Figure 30, Location 9), rocks of the Stoughton - Roquemaure Group envelop an oval area of felsic rocks belonging to the Hunter Mine Group. No contacts are exposed, but in 2 places, fine-grained tholeiitic basalt dikes cut the rhyolites. By comparison with the base of the type section and the section in South Bay, this area probably represents a knoll of felsic volcanic rock which was engulfed by mafic lava flows of the Stoughton - Roquemaure Group.

The gabbroic textured rocks are similar to the gabbroic rocks at the base of the Stoughton - Roquemaure Group in Roquemaure Township, Quebec. On the Ol′-Ne′-Q′, An-Ab′-Or, and Jensen Cation Diagram (Figures 35, 36, and 37), the coarse- grained rocks project as subalkalic, orthoclase-poor magnesium-rich to iron-rich tholeiitic basalts as do the chemical analyses (Eakins 1972) of the gabbroic rocks from Roquemaure Township shown on the same diagrams. The magnesium-rich tholeiitic basalt-types are green, equigranular rocks having a grain size of 2 to 3

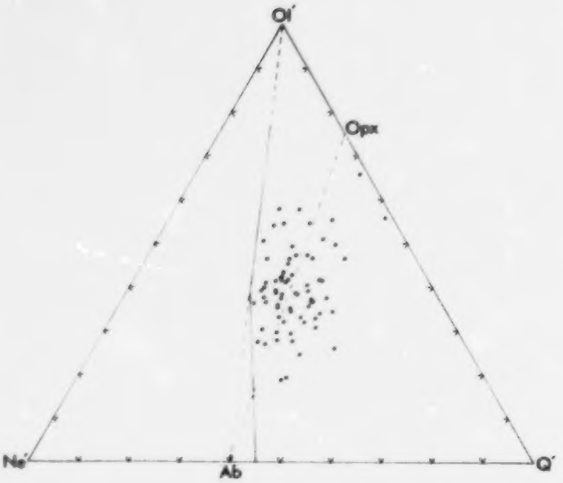


Figure 38. Ol'-Ab'-Q' projections of analyses of the upper volcanic flows of the Stoughton - Roquemaure Group are from the western part of the Lake Abitibi section.

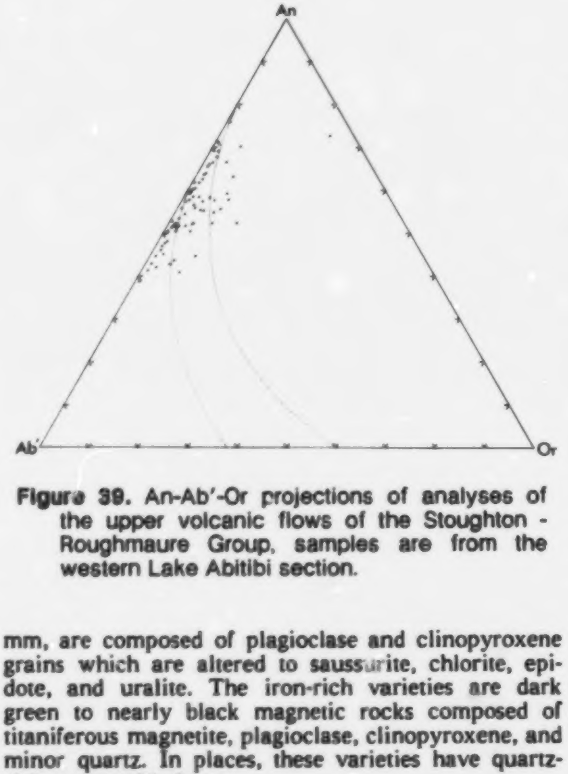


Figure 39. An-Ab'-Or projections of analyses of the upper volcanic flows of the Stoughton - Roughmaure Group, samples are from the western Lake Abitibi section.

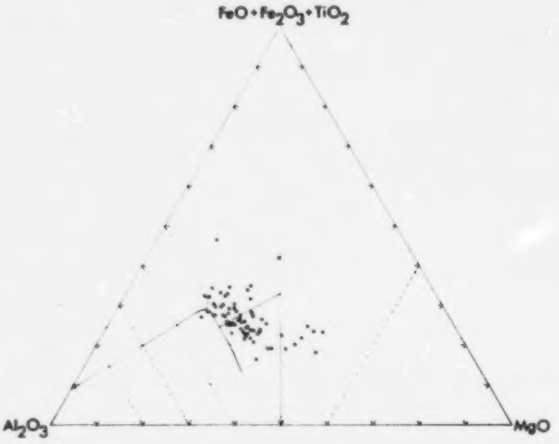


Figure 40. Projections of analyses of the upper volcanic flows of the Stoughton - Roquemaure Group from the western Lake Abitibi section on the Jensen Cation Diagram.

mm, are composed of plagioclase and clinopyroxene grains which are altered to saussurite, chlorite, epidote, and uralite. The iron-rich varieties are dark green to nearly black magnetic rocks composed of titaniferous magnetite, plagioclase, clinopyroxene, and minor quartz. In places, these varieties have quartz-rich pegmatoid phases.

The coarse-grained rocks form units 50 to 100 m thick that alternate with fine grained units 2 to 5 m thick. The fine-grained map units strike straight across outcrops and can be followed for 2 to 3 km from outcrop to outcrop. They contain no recognizable, chilled contacts. Similar "sheeted" gabbroic rocks have been described in the Labrador Trough by Dimroth *et al.* (1970). A few of the fine-grained map units represent a gradational, compositional change in the rocks from magnesium-rich to iron-rich tholeiite and some of the coarse-grained magnesium-rich tholeiites appear to grade laterally into fine-grained pillowed lavas of the same composition. These rocks are believed to represent lava flows and therefore, have been shown as volcanic rocks on (Map P.2433, back pocket).

The coarse-grained tholeiites maybe lava flows. These rocks may represent either rocks correlative with the coarse-grained rocks found near the base of the type section or may represent thick lava flows which accumulated in depressions to produce a smoother topography and to allow subsequent lava flows to spread out and be thinner. The coarse-grained rocks do not show mineralogical layering.

Peridotitic intrusions were not found in the area nor were peridotitic komatiites which may indicate that either these rocks were not formed in the area, or that they were extruded at lower elevations on the slopes of the calc-alkalic volcanic pile. These volcanic sequences show a general upward decrease in the amounts of komatiitic lavas present. Thus, the base of the Stoughton -Roquemaure Group, then could represent mafic lavas at a higher elevation around a felsic calc-alkalic volcanic pile than those lavas at the base of the type section in Roquemaure Township. As was pointed out earlier in the report, rocks of the western Lake Abitibi section of the Hunter Mine Group were probably deposited at higher elevations than were rocks of the Hunter Mine Group in Roquemaure Township. These similar observations could be related or could be coincidental.

The finer grained lavas which overlie the coarse-grained rocks, appear to be dominantly massive and pillowed magnesium tholeiitic lavas intercalated with lesser amounts of basaltic komatiite and iron-rich tholeiitic basalt. Analyses of these lavas have been projected on Ol'-Ne-Q, An-Ab'-Or, and on Jensen Cation diagrams (Figures 38, 39, and 40). Like the

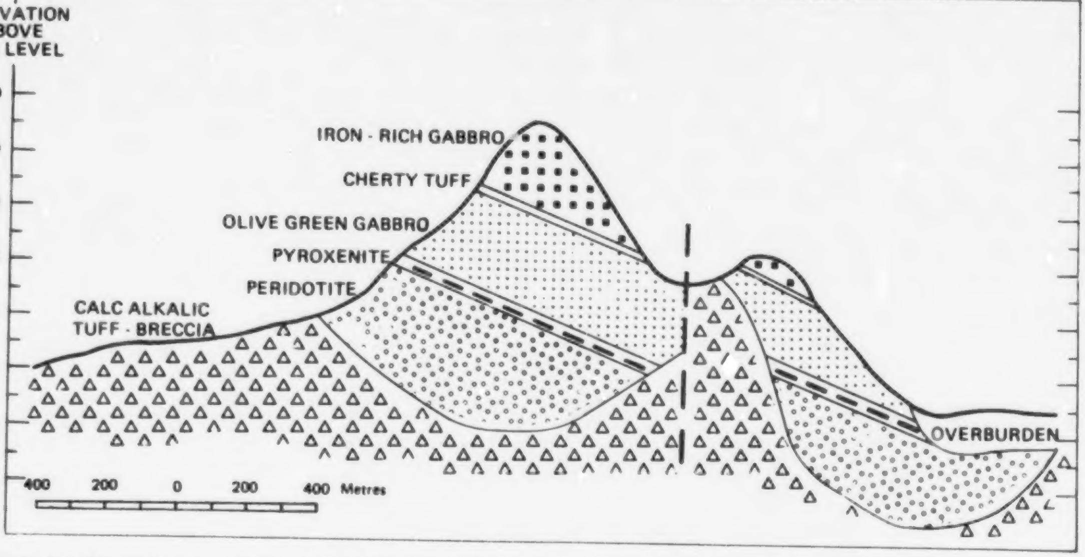


Figure 41. Cross-section of the Ghost Range sections. Vertical exaggeration is 5.3X horizontal.

rocks of the type section of the Stoughton - Roquemaure Group, they are subalkalic and orthoclase-poor basaltic komatiites and tholeiites. In spite of the limited amount of outcrop, it seems unlikely that such peridotitic komatiites are present in this section. The presence of polyhedrally-jointed basaltic komatiites in this section in conjunction with the other geological data, however, confirms that this volcanic succession is part of the Stoughton - Roquemaure Group. On the Jensen Cation Diagram (Figure 40), the one high point in the field of iron-rich tholeiitic basalt represents a rock analysis from the thick iron-rich tholeiitic basalt flow at the top of the Stoughton -Roquemaure Group.

Ghost Range Section

The Ghost Range Section occurs in the southern parts of Lamplugh and Frecheville Townships and in the adjoining northern parts of Harker and Holloway Townships (Figure 30, Location 12). The section has been mapped by Satterly (1952b, 1954) as layered ultramafic to mafic sills which were subsequently folded into a doubly plunging syncline. Later, MacRae (1969) concurred with Satterly's conclusions and called it the Ghost Range Sill which he described as a simple, 760 m thick, layered sequence of peridotite - clinopyroxenite - gabbro.

Instead of being a folded sill, the Ghost Range Section probably represents a basal sequence of the Stoughton - Roquemaure Group, and is composed of thick, flat-lying lava flows, resting unconformably on folded calc-alkalic volcanic rocks of the Hunter Mine Group. The Ghost Range is a 10 km long ridge with steep, 200 m relief. Calc-alkalic rocks of the Hunter Mine Group are exposed on all sides of the ridge and in some deep valleys within it (Figure 41). The contacts between the rock units within the Ghost Range Section have dip less than 5° to the northeast.

The contact of the peridotite with lower calc-alkalic volcanic rocks is exposed only at the base of Lightning Mountain (Satterly 1954; Figure 30, Location 13), and is considered to be a vertical fault contact. Sheared peridotite rests against pillowed lavas of calc-alkalic basalt. Diamond-drill hole intersections of peridotite 100 m below the base of the Lightning Mountain Ridge, suggest that the peridotite is up to 140 m thick; but this thickness may be restricted to valleys in the paleotopography on the calc-alkalic volcanic rocks. In places, knolls of calc-alkalic volcanic rocks appear to have been high enough not to have been covered by peridotite (Figure 41).

The rock units in the Ghost Range Section in ascending order, are peridotite, clinopyroxenite, olive-green gabbro, a thin layer of cherty tuff, and a dark green to black iron-rich gabbro (Figure 41). Chemical analyses of these rocks have Ol'-Ne'-Q'and An-Ab'-Or projections (Figures 42 and 43) which show all of them to be mainly subalkalic and orthoclase-poor. On the Jensen Cation Diagram (Figure 44), the peridotites lie in the peridotitic komatiite field, and the clinopyroxenites and olive-green gabbros lie in the fields of basaltic komatiite, magnesium-rich tholeiite, and calc- alkalic basalt. The upper, dark green to black gabbros occur in the iron-rich tholeiitic basalt and tholeiitic andesite fields as a cluster of points quite separate from those which represent the lower rocks. This suggests that the 2 types of gabbros are distinct (Figure 44).

MacRae (1969) interpreted the peridotite and pyroxenite to have resulted from olivine and pyroxene crystals settling in a magma chamber. The lower peridotite contains zones of dunite (Satterly 1952a, 1952b, 1954) and rodingite (MacRae 1969).

MacRae (1969) described a 2 to 6 m thick clinopyroxenite, intermediate to both the gabbro and peridotite, as being intensely sheared over a width of a few cm at its lower contact with the peridotite, and

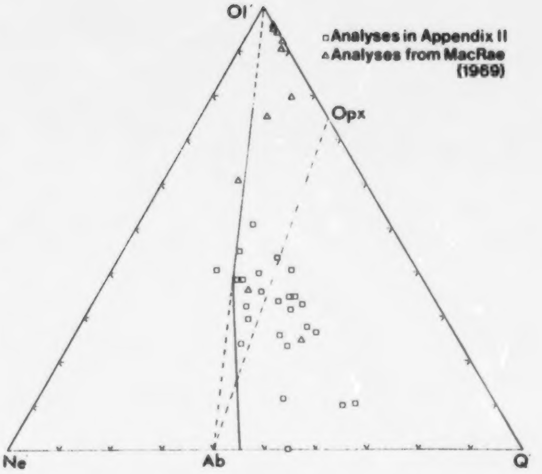


Figure 42. Ol'-Ab'-Q'projections of chemical analyses from the Ghost Range and Garrison sections. The open squares represent analyses in Appendix II and the open triangles represent analyses from MacRae (1969).

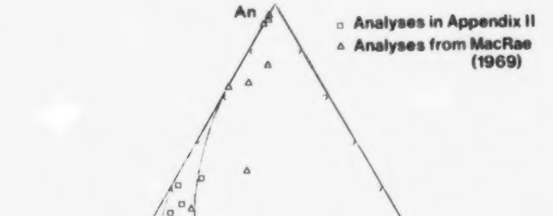


Figure 43. An-Ab'-Or projections of chemical analyses from the Ghost Range and Garrison sections. The open squares represent analyses in Appendix II and open triangles represent analyses from MacRae (1969).

also as having a sharp contact with the overlying gabbro; there was no evidence of chilling. Where the lower contact was re-examined, no sign of shearing was found. Instead, a gradational contact about 10 cm thick was observed. The upper contact between the clinopyroxenite and gabbro was found to be sharp with a slight decrease in grain size in the gabbro.

The gabbro above the clinopyroxenite has a grain size of 2 to 3 mm is a light olive-green rock, and is composed of 40 to 60% clinopyroxene, 40 to 60% plagioclase (An_{60} to An_{80}), and accessory serpentine and magnetite pseudomorphic after olivine. The gabbro is about 120 m thick and ranges in chemical composition from basaltic komatiite to magnesium-rich tholeiitic basalt, and, where plagioclase rich, to calc-alkalic basalt (Figure 44).

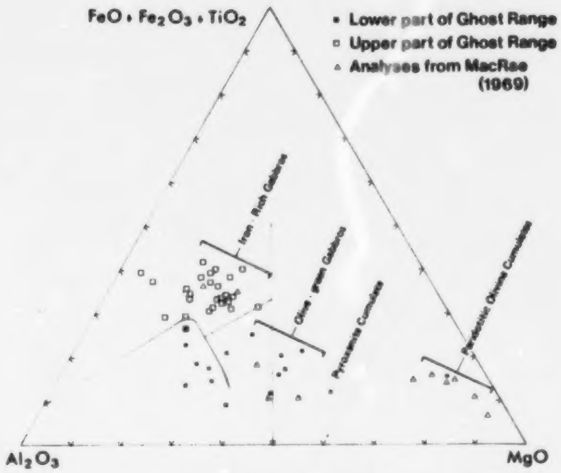


Figure 44. Jensen Cation Diagram projections of chemical analyses of Ghost Range section and Garrison section. Solid squares and open squares are from the lower and upper parts of the sections, respectively, and the open triangles are from MacRae (1969).

Bedded calc-alkalic felsic tuff is 10 cm to 3 m thick, has dips of about 10°, and overlies the olive-green gabbro. The tuff appears to have been deposited on the olive-green gabbro prior to the deposition of the overlying iron-rich gabbro. The gabbroic rocks both above and below the tuff, become fine grained toward their contacts with the tuff, but it was not possible to determine from the contacts whether the gabbroic rocks were extrusive or intrusive.

The gabbroic rocks which form the upper unit in the Stoughton - Roquemaure section above the tuff are dark green to black magnetic rocks, have a grain size of 2 to 4 mm, and have diabasic textures. These rocks contain 40 to 50% pyroxene, 30 to 40% plagioclase (An_{40} to An_{60}), 2 to 10% quartz, and 4 to 8% titaniferous magnetite. The iron-rich parts have numerous irregular zones of pegmatitic granophyre similar to those observed in the iron-rich gabbroic rocks at the western end of Lake Abitibi and in Roquemaure Township. The iron-rich gabbroic rocks form layers 20 to 50 m thick separated by fine-grained layers 3 to 10 m thick. The fine-grained rocks grade upward into the coarse-grained rocks and some extend for 1.5 to 3 km which indicates that they may represent either flow or intrusive contacts. Although no chilled contacts were detected, the author considers the iron-rich gabbros to represent massive tabular flows of iron-rich tholeiitic basalt.

Fine-grained layers were not observed in the olive-green gabbros below the tuffs. The olive-green gabbro, as well as the peridotite and pyroxenite were possibly derived from the same magma. This magma may have represented a komatiitic lava flow in which olivine and pyroxene crystals settled. After solidification, the flow was overlain by tuff and additional flows of iron-rich tholeiitic basalt.

MacRae (1969) had concluded that all the gabbros in the Ghost Range Section were derived from the fractionation of olivine and clinopyroxene. Arndt (1975) used Theo's Flow specifically to illustrate that iron-rich gabbro can form by the fractionation of olivine and pyroxene from a tholeiitic picritic magma. In Theo's Flow, the Al_2O_3/TiO_2 ratios of rock types are about 10, while in Fred's Flow, which Arndt (1975) considered to be a komatiitic magma, the ratios are about 20 (Figure 45). By comparison to these flows, the rock in the Ghost Range Section, if they were formed from 1 magma, should have a constant Al_2O_3/TiO_2 ratio. Instead, the lower rocks (peridotite, pyroxenite, and olive-green gabbro) have ratios of about 20 and the iron-rich gabbro, about 10, which suggests that iron-rich gabbros are unrelated to the lower rocks, and that the lower rocks were derived from a komatiitic magma. If the lower rocks represented a flow, the flow would have had to be about 400 m thick or about 3 times as thick as Fred's Flow.

A layer of komatiitic magma greater than 400 m thick would indicate that komatiitic lavas are more abundant in the Ghost Range Section than in the type section of the Stoughton - Roquemaure Group. As a lava flow, this thick komatiite must have erupted and accumulated on the edges of the calc-alkalic volcanic pile; or, it may have been blocked from spreading into the type section. The iron-rich gabbroic rocks above the komatiite are stratigraphically correlative with, or lower than the gabbroic rocks at the base of the Stoughton - Roquemaure type section.

Since the Stoughton - Roquemaure Group in the Ghost Range Section is a flat-lying section, with little deformation or metamorphism, the thick accumulations of lavas in the type area could not have been deposited on top of the presently preserved Ghost Range Section. This suggests that the Ghost Range Section remained high between the northern and middle branches of the Destor - Porcupine Fault Zone, and was buoyed up by the underlying felsic volcanic rocks. The upper parts of the type section and the western Lake Abitibi sections accumulated in a sinking rift zone that extended into the calc-alkalic pile as a branch fault of the developing Destor - Porcupine Fault Zone to the south (*see* Figures 5 and 30).

Garrison Section

The Garrison Section (*see* Figure 30, Location 15) of the Stoughton - Roquemaure Group dips and faces 60 to 70° north. It is a 760 m thick, layered, peridotite - pyroxenite - gabbro sill, overlain by lava flows (Satterly 1949b; MacRae 1969). Only the upper 400 m of the section is exposed. Geophysical exploration and diamond drilling indicate that the lower peridotite overlies cherty tuff and tuff-breccia intercalated with jasper iron formation belonging to the Hunter Mine Group. The peridotite is extensively serpentinized and is overlain by a 10 m thick layer of pyroxenite, which is in turn overlain by a layer of gabbro about 100 m thick. The gabbro is chemically equivalent to basaltic komatiite and magnesium-rich tholeiitic basalt. The contact between the gabbro layer and the overlying iron-rich tholeiitic basalts is covered, and thus, its nature is unknown. The iron-rich tholeiitic basalts are 200 m thick and (at the top) are in fault contact with the older calc-alkalic rocks of the Hunter Mine Group.

MacRae (1969) pointed out that the proportion of peridotite to gabbro in the Garrison Section was much greater than that found elsewhere. He suggested either that the parent magma carried a high content of suspended olivine crystals during its emplacement, or that the residual magma had escaped after much of the olivine and pyroxene had separated. The high proportion of peridotite to gabbro could also have been caused by thickening of the peridotite or a removal of part of the gabbro by faulting.

The Al_2O_3/TiO_2 ratios (Figure 45) from analyses made on the peridotite and gabbros of the Garrison Section suggest by comparison with Fred's Flow (Arndt 1975), that the rocks were derived from a komatiitic magma. If the present thicknesses of the rocks are the original ones, the magma must have had a magnesium content of about 30%.

The peridotite and gabbro may have been a lava flow or a layered sill. The presence of the Hunter Mine Group at the base of the peridotite and iron-rich tholeiitic basalt lavas above the gabbro suggest that these rocks formed from a pool of lava on the paleosurface of the Hunter Mine Group. The iron formation in the Hunter Mine Group suggests that the Garrison Section of the Stoughton - Roquemaure Group was deposited at a lower elevation on the slope of the pre-existing calc-alkalic pile than were the rocks of the Ghost Range Section which is underlain by Hunter Mine Group rocks devoid of iron formation and other sedimentary rocks. In addition, the tilting of the rocks to the north suggests that the rocks of the Garrison Section were more strongly affected by activity along the Destor - Porcupine Fault Zone than were the rocks of the Ghost Range Section.

A few isolated outcrops of peridotitic rocks occur near the Garrison Section (*see* Figure 30, Location 14). These appear to be non-layered bodies in the form of stocks and sills (Satterly 1949a, 1949b), which intruded the calc-alkalic rocks of the Hunter Mine Group.

Destor - Porcupine Section

The Destor - Porcupine Section of the Stoughton - Roquemaure Group is located between the middle and southern branches of the Destor - Porcupine Fault Zone (*see* Figure 30, Location 16 to 17). Here, the rocks are sliced into numerous wedges and are intruded by syenitic rocks which have extensively altered the rocks in places. Almost all the information about this section has been gleaned from the diamond drilling logs.

Along the middle branch of the Destor - Porcupine Fault Zone, wedges of komatiitic rock appear to have been juxtaposed with calc-alkalic rocks of the Hunter Mine Group on the northern side of the fault. Along the southern branch of the Destor - Porcupine Fault Zone, the komatiitic rocks are juxtaposed by faulting with younger south-facing tholeiitic lavas of the Kinojevis Group. Between the 2 main faults, the komatiitic rocks are intercalated with wedges of sedi-

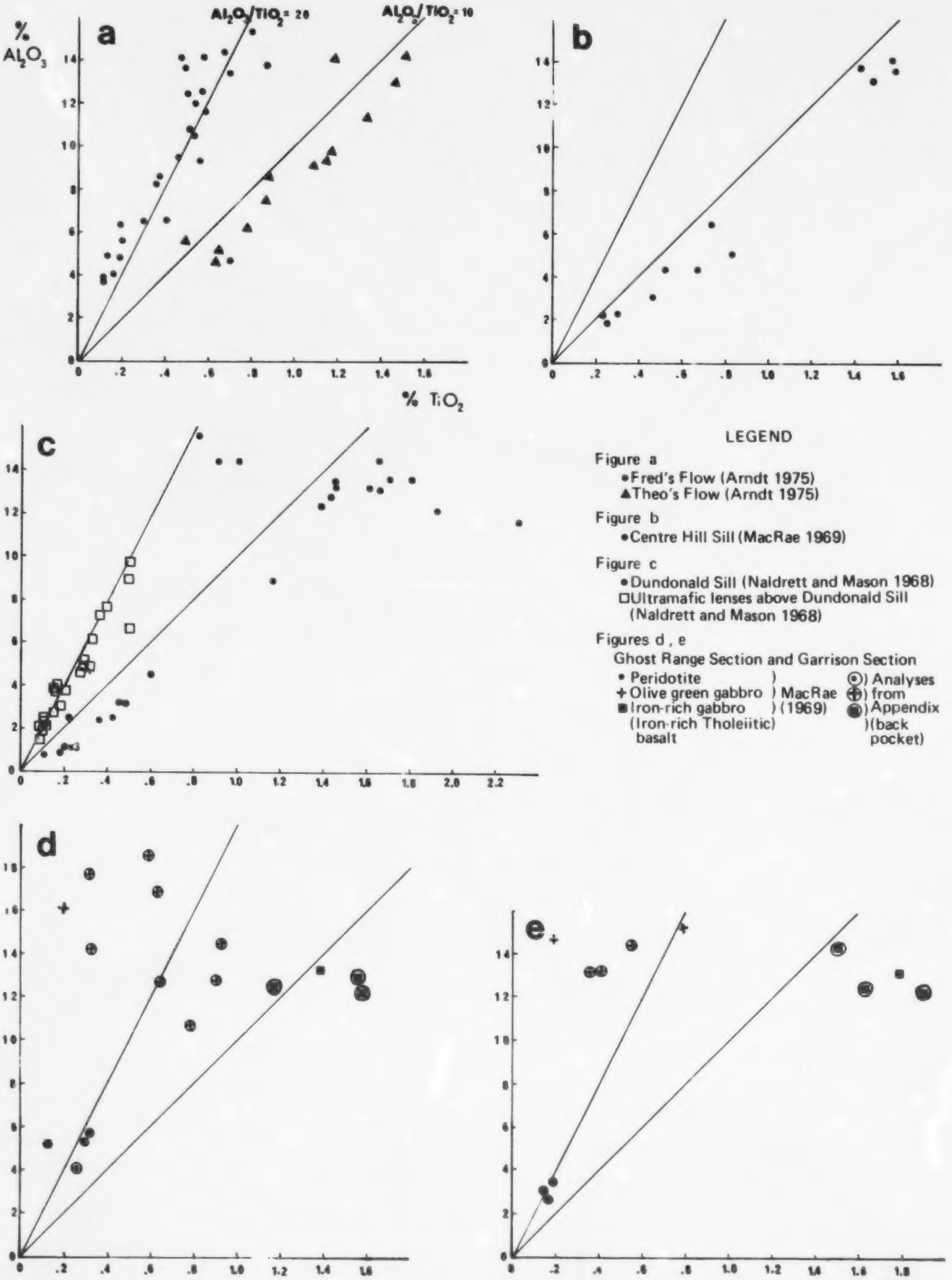


Figure 45. Percentage of Al_2O_3 versus percentage of TiO_2 in flows and sills in the Timmins - Kirkland Lake area north of the Destor - Porcupine Fault zone.

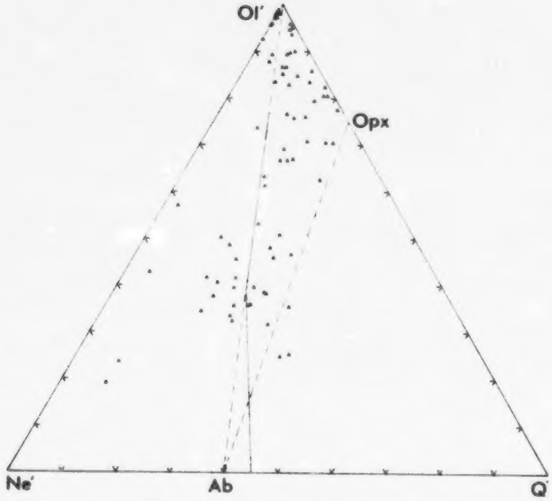


Figure 46. Ol'-Ne'-Q' projections of analyses of Stoughton -Roquemaure Group along the Destor - Porcupine Fault zone.

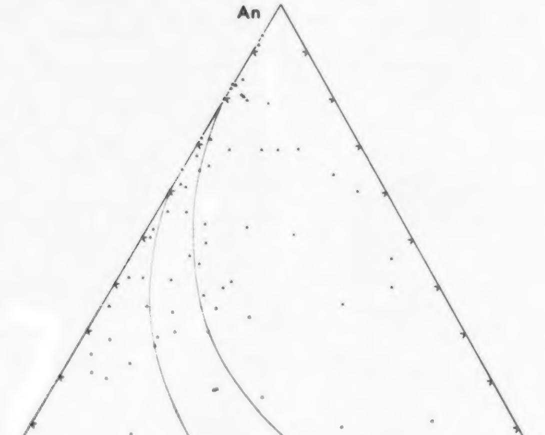


Figure 47. An-Ab'-Or projections of analyses of the Stoughton - Roquemaure Group along the Destor - Porcupine Fault zone.

mentary rock and alkalic volcanic rock. The faults are shear zones composed mainly of chlorite, talc, serpentine, and actinolite.

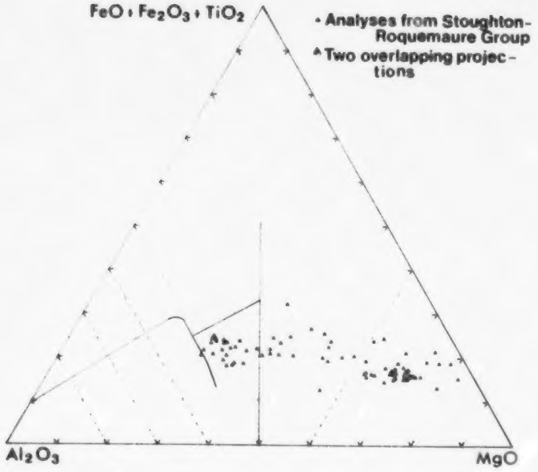


Figure 48. Jensen Cation plots of analyses of the Stoughton - Roquemaure Group along the Destor - Porcupine Fault zone (large black triangles equal 2 overlapping projections).

The Ol'-Ne'-Q'diagram (Figure 46) shows that about a third of rocks from this area represent alkalic rocks. These rocks are presumed to have been enriched by alkalies derived from numerous alkalic intrusions along the southern part of the Destor - Porcupine Fault Zone. The An-Ab'-Or projections of the rocks (Figure 47) are scattered over the whole diagram to show that K-enrichment as well as Na-enrichment has occurred. However, on the Jensen Cation Diagram (Figure 48), the rocks follow a continuum from the peridotitic komatiite field to the magnesium-rich tholeiitic basalt field. In spite of the strong alkali and carbonate alteration of some rocks, the analyses of these rocks still fall within the rock-type fields to which they would have been accorded, based on thin section and field examinations. In many places, the polyhedral jointing and relict olivine spinifex textures are preserved in the strongly altered peridotitic komatiites, and many of the rocks have chrome-mica, serpentine, talc, magnesite, and actinolite, to indicate their original compositions.

In the Destor - Porcupine Section, the proportion of peridotitic komatiite to basaltic komatiite and magnesium-rich tholeiitic basalt appears to be greater than that found in the Western Lake Abitibi Section and in type sections of the Stoughton - Roquemaure Group. Because peridotitic komatiite appears to be more abundant toward the base of a komatiitic succession, this portion of the Stoughton - Roquemaure Group may represent a lower section of the group developed in a basin marginal to the calc-alkalic pile. This indicates that the Stoughton - Roquemaure Group thickened southward, unless the upper parts of the group were not formed in this area. However, south of the southern branch of the Destor - Porcupine Fault Zone (*see* Figure 5), the upper equivalents of the type section can be recognized farther west (*see* section on Lower Formation, Tisdale Group) indicating that the Stoughton - Roquemaure Group was at least 10 km thick along the Destor - Porcupine Fault Zone.

The presence of the Kinojevis Group south of the fault suggests that the upper part of the Stoughton - Roquemaure Group was downfaulted to the south for at least 10 km explaining the development of the talc-chlorite schists found in abundance here. In addition, the carbonatized komatiites here, probably resulted from alkali and CO_2-bearing hydrous solutions that accompanied the alkalic intrusions penetrating upward along deep fractures. Rocks of the Stoughton - Roquemaure Group in the Ghost Range Section north of the middle branch of the Destor - Porcupine Fault have not been so affected (*see* Figures 42 and 43).

THE MUNRO KOMATIITIC SUCCESSION

Several layered bodies of ultramafic to mafic rock occur in Munro, and adjoining McCool and Beatty Townships west of the Lightning River area (Figure 49) and north of the Destor - Porcupine Fault Zone (Satterly and Armstrong 1949; Satterly 1952a, 1952b; MacRae 1969; Arndt 1975; and Arndt *et al.* 1977). Arndt (1975) showed that 2 of these bodies of rock were thick lava extrusions: 'Fred's Flow' being a komatiitic lava flow, and 'Theo's Flow', a picritic tholeiitic lava flow.

In the volcanic succession of northern Munro Township Arndt (1975) found that the komatiitic lavas were underlain by tholeiitic lavas and tholeiitic layered peridotite-gabbro sills. He suggested that the tholeiitic rocks represented early partial melts of a rising mantle diapir. The coarse-grained rocks which he described as being tholeiitic appear to be underlain by felsic volcanic rocks to the north in Warden and northern McCool Townships (Satterly 1953). These felsic volcanic rocks extend into the Lightning River area and are considered to be part of the Hunter Mine Group (*see* Figure 30). This suggests that komatiites and their underlying coarse-grained tholeiitic rocks may be equivalent to other basal sections of the Stoughton - Roquemaure Group.

The gabbros in the peridotite-gabbro bodies described as sills in Munro Township by MacRae (1969) are iron-rich tholeiitic gabbros underlain by layers of pyroxenite and olivine crystals (Arndt 1975). The basal peridotites and pyroxenites and their overlying iron-rich gabbros have Al_2O_3/TiO_2 ratios of about 10, which implies that they formed in a similar manner to Theo's Flow (*see* Figure 45). The iron-rich tholeiitic gabbroic rocks in the Lightning River part of the Stoughton - Roquemaure Group do not have basal olivine cumulate layers and may represent thick tholeiitic lavas that were tapped from magma chambers, in a similar manner to the formation of sills in Munro and McCool Townships. The iron-rich gabbros of Munro Township have pegmatoid and aphanitic zones (Arndt 1975) like the iron-rich gabbroic rocks at the base of the Stoughton - Roquemaure type section, and the upper part of the Ghost Range Section.

By analogy with the Lightning River area, it would appear that the komatiitic and tholeiitic rocks in Munro Township represented the flooding of a calc-alkalic volcanic pile by mafic to ultramafic lavas. Additional komatiitic and tholeiitic lavas probably were extruded to the south at lower elevations on the paleosurface of the calc-alkalic pile.

LOWER FORMATION OF THE TISDALE GROUP

In the Timmins area, Pyke (1980) used the term "Lower Formation of the Tisdale Group" to refer to the komatiitic succession (*see* Figure 5) which he considered to be equivalent to the Stoughton - Roquemaure Group in the Kirkland Lake area. In the Lower Formation, Pyke (1980) included the Bowman Subgroup of Goodwin (1977a, 1977b) and the komatiitic rocks of Dundonald Township (Naldrett and Mason 1968; Figure 49).

North of the Destor - Porcupine Fault Zone, in Dundonald Township, the komatiitic and gabbroic rocks described by Naldrett and Mason (1968) appear to overlie felsic volcanic rocks (Muir and Comba 1979). North of Timmins (Figure 49), the komatiites appear to overlie as well as interfinger with the sedimentary rocks of the Porcupine Group (Pyke 1980) and all appear to be a westward continuation of the komatiitic rocks in Beatty, Munro, and McCool Townships.

South of the Destor - Porcupine Fault Zone the Lower Formation of the Tisdale Group begins north of the Watabeag Batholith (*see* Figure 99) where these rocks have been referred to as the Bowman Subgroup of the Blake River Group by Goodwin (1977a, 1977b; Figures 5 and 49). Here, the succession is dominantly composed of basaltic komatiites and magnesium-rich tholeiitic basalts. To the north, its base is truncated by the southern branch of the Destor - Porcupine Fault Zone, and to the south, its top consists of a diabasic, iron-rich tholeiitic basalt flow which is overlain by calc-alkalic rhyolite tuff-breccia, cherty tuff, and weakly developed cherty iron formation. This zone of basalt - rhyolite - iron formation is similar to that found at the top of the Stoughton - Roquemaure Group at its type section, except that here the zone is thicker. As in the type section, komatiitic lavas do not occur in the strata above the iron-rich tholeiitic basalt. The komatiitic succession in Bowman Township indicates that komatiitic volcanism occurred south as well as north of the Destor - Porcupine Fault Zone.

South of Timmins, the Lower Formation of the Tisdale Group appears to conformably overlie the felsic volcanic rocks and the iron formation of the Upper Formation of the Deloro Group (Pyke 1980; Figure 5). In many places, the upper parts of the Upper Formation of the Deloro Group are intruded by peridotite stocks which may have been feeders for the komatiitic lavas (Pyke 1982).

From Timmins, the Lower Formation of the Tisdale Group extends south and then east, where it becomes truncated by the Kirkland - Larder Lake Fault Zone (*see* Figure 5). Throughout this area, the komatiitic lavas of the Lower Formation are steeply dipping and face mainly toward the centre of the synclinorium. In most places, the komatiitic succession is less than 5 km thick, which would indicate that the komatiitic succession at the base of the Upper Supergroup became thinner to the west.

By analogy with the Stoughton - Roquemaure Group, the Lower Formation of the Tisdale Group probably represented a westward encroachment of komatiitic and tholeiitic lavas over the lower parts of earlier formed felsic volcanic piles. As the low areas were filled up, the komatiites were able to lap onto the felsic volcanic rocks where they are now located. Deeper in the basin to the east, the komatiitic lavas must have interfingered with and covered the earlier formed sedimentary rocks (*see* Figure 29).

LARDER LAKE GROUP

Area and History

The Larder Lake Group is a komatiitic succession that begins south of the Kirkland Lake - Larder Lake Fault Zone, that is south of the Watabeag Batholith (*see*

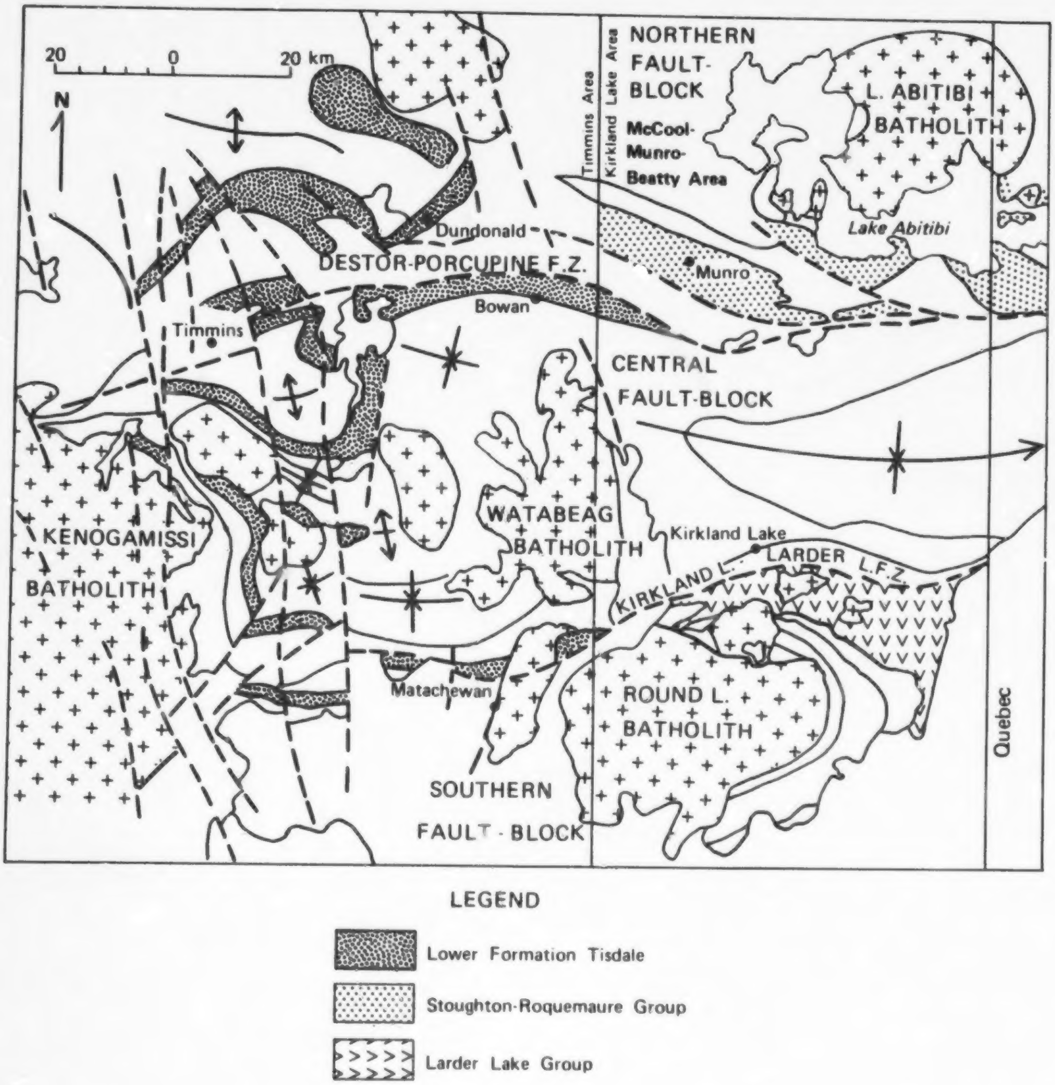


Figure 49. Geological map of the Timmins - Kirkland Lake area showing the distribution of komatiitic volcanic successions.

Figure 5) and widens eastward, as it approaches the Ontario - Quebec boundary where it disappears under Proterozoic rocks (Figure 50). The group includes calc-alkalic rhyolite tuff-breccias and tuffs, clastic sedimentary rocks, iron formation, limestone, and dolomite as well as komatiitic and tholeiitic lavas (Jensen 1978c), all of which have been folded nearly isoclinally.

Ridler (1970) used the name "McVittie Basalts" when referring to the mafic lava flows south of the Kirkland Lake - Larder Lake Fault Zone because he correlated the mafic lava flows of McVittie Township north of the fault with those to the south. Subsequently, Jensen (1978c) found that the mafic lavas in McVittie Township actually belong to the Kinojevis Group because these lavas are a tholeiitic succession with no komatiitic lavas, rhyolites, and sedimentary rocks, unlike the "McVittie Basalts" south of the fault. Consequently, Jensen (1978c) suggested that these rocks south of the fault be called the Larder Lake Group. Jensen (1978c) also showed that south of the Kirkland Lake - Larder Lake Fault Zone, the Timiskaming sedimentary rocks of Thomson (1949) and Hewitt (1951) were interlayered with the komatiitic lavas and rhyolite tuffs and were distinct from the Timiskaming sedimentary rocks near Kirkland Lake, as had been suggested by Kalliokoski (1968). Jensen (1978c) also concluded that the komatiitic lavas extended west and included the volcanic rocks considered by Ridler (1970) to be Timiskaming, as well as the volcanic rocks Ridler considered to be younger than the Timiskaming and which he called the "Highway 11 Basalts".

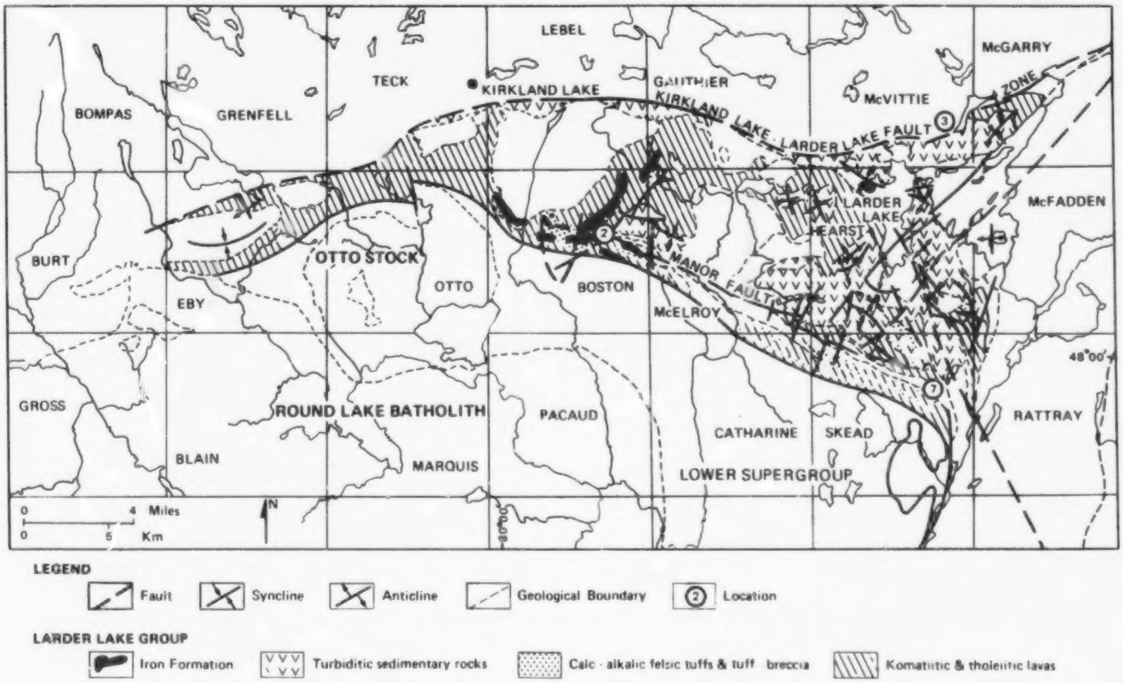


Figure 50. Map showing the distribution of the rock types of the Larder Lake Group.

Stratigraphy

The stratigraphy of the Larder Lake Group suggests that the group was formed on a broad shelf between an elevated area to the south and a marine basin to the north and that the shelf may have had an irregular topography. Although the komatiitic and tholeiitic lavas are uniform, the rhyolite tuffs and sedimentary rocks reveal substantial facies changes throughout the area. The thickness of the Larder Lake Group is estimated to be about 5 to 6 km and the units are younger to the northwest. As well, the sequence is overlain by tholeiitic lavas considered to be part of the Kinojevis Group southwest of Kirkland Lake (*see* Figure 5).

The lower contact of the Larder Lake Group with the Skead Group has yet to be comprehensively studied in depth. Hewitt (1951) implied that there was a conformable contact between the 2 groups of rock and he mentioned that the overlying mafic flows were interlayered by acid fragmental rocks which were similar to those found in the Skead Group. These tuffs and tuff-breccias have been examined by the author and found to be the most prominent in the basal part of the more easterly parts of the Larder Lake Group (*see* Figure 50, Location 1). To the west and north, the horizons of tuff and tuff-breccia are thinner and finer grained, where they can be described as cherty tuffs. Westward, in the northern part of Boston Township, the horizons of tuff and tuff-breccia are interbedded with the iron formation (the Timiskaming Boston Iron Formation of Ridler (1970); *see* Figure 50, Location 2). The presence of iron formation suggests that the tuffs were deposited in deep quiet water, away from their source area to the southeast (Dimroth 1975b). The komatiitic and tholeiitic lavas form extensive layering of tabular and pillowed flows that show little or no change near the base of the Larder Lake Group as do the tuffs and tuff-breccias. The komatiitic lava flows contain thick zones of olivine spinifex suggesting (Arndt *et al.* 1977) that they were derived from local sources throughout the area of deposition. Numerous circular intrusions of peridotite cut the rocks near the base of the Larder Lake Group, and may have been the feeders.

Rocks at the base of the Larder Lake Group suggest that there was an overlapping of felsic volcanism from a central vent area to the southeast, and pervasive komatiitic and tholeiitic volcanism from local sources. Calc-alkalic andesites and basalts are absent in the sequence, and the lower part of the Larder Lake Group has a bimodal volcanic suite like those commonly found near the base of many other greenstone belts (Glikson 1976c; Glikson and Lambert 1976; Barker and Peterman 1974; Shaw 1976; Hubregtse 1976).

In the central parts of the Larder Lake Group, the komatiitic lavas, tholeiitic lavas, and felsic tuffs become interlayered with turbiditic sedimentary rocks composed of conglomerates, greywackes, and argillaceous rocks (Figure 50). Hyde (1978) gave a brief description of these rocks.

The author found the sedimentary rocks to be well preserved and showed all the textures and facies of turbidites (Walker 1976). The dominant conglomerate is chaotic, matrix supported, and has unsorted sand to boulder-sized volcanic clasts of peridotitic komatiite, basaltic komatiite, magnesium-rich tholeiitic basalt, and quartz-porphyry rhyolite tuff-breccia set in a talc-chlorite matrix. Rhyolite clasts up to 10 m across occur in some conglomerates. The proportion of rhyolite clasts can vary from 0 to 60% in separate beds of conglomerate. These conglomerates suggest that a steep scarp existed to the south.

The other types of conglomerate are clast-supported conglomerates with imbrication and normal and reverse grading of the clasts. Many of these clast-supported conglomerates, in which rhyolite clasts are absent, are difficult to distinguish from polyhedrally jointed, weakly deformed peridotitic komatiites. These conglomerates can be recognized only where isolated olivine spinifex textured clasts are present. As a consequence, many of the rocks have been previously mapped as flows.

The greywackes form beds 5 to 100 cm thick, in which complete and partial Bouma cycles can be recognized. Many beds show deep channelling into the underlying beds typical of turbidites. The argillaceous rocks form graded beds 1 to 2 cm thick, many of them containing flame structures and load casts. Slump folds are common among groups of beds.

The turbidites extend from the southwest up to the edge of the Kirkland Lake - Larder Lake Fault Zone, where they were juxtaposed by faulting with the Timiskaming sedimentary rocks (*see* Figure 50, Location 3). Near the fault zone, the argillaceous rocks contain layers of iron formation (magnetite) and in places, they grade into limestone and dolomite.

The sedimentary material appears to have been derived from the south, and, most of the coarse material must have been deposited at the foot of a volcanic pile (scarp) on a shelf that extended north from Skead Township to the Kirkland Lake - Larder Lake Fault Zone (*see* Figure 50). Also, the komatiitic and tholeiitic lavas remain abundant throughout the area.

Structure

The Larder Lake Group, particularly in its eastern part, is isoclinally folded. Except along narrow shear zones, the rocks show little or no penetrative deformation to suggest that they were folded at depth by confining pressures. The isoclinal folding does not extend into the lower part of the Skead Group which as was mentioned earlier, appears to represent a volcanic pile turned on its side. Metamorphism in the basal rocks of this pile does not suggest burial under 9 km of rock.

The difference in folding between the Skead and Larder Lake Groups suggests that the eastern side of the volcanic pile sank as the pile formed (*see* Figure 50), and that the western side was buoyed up by the earlier developed volcanic pile represented by the Pacaud Tuffs. Because the lavas of the Skead Group would have accumulated in greater volume to the east, their greater weight would have increased the tilting of the volcanic pile eastward. In other words, the volcanic pile would have been pulled downwards to the east as it formed. This process would have been greatly facilitated if the rocks at the base of the pile to the east were transformed into eclogite, thus decreasing the underlying volume of rock. The heavier eclogite would also sink. Since the formation and partial melting of eclogite was probably required to produce the rhyolitic rocks of the Skead and Larder Lake Groups (Ewing 1979), the process of subsidence following the sinking eclogite becomes more plausible. The isoclinal folding without confining pressure to cause deformation seen in the Larder Lake Group was also probably caused by the formation of eclogite at depth which would have caused the shelf to buckle and pull the Larder Lake Group into isoclinal troughs. This idea will be discussed further in this report.

As the Skead monoclinal volcanic sequence turned downward and the Larder Lake Group remained buoyant, a fault zone should have formed between the 2 areas. This fault is probably represented by the Manor Fault (*see* Figure 50) recognized by Hewitt (1951).

WABEWAWA GROUP

The Wabewawa Group represents a komatiitic volcanic succession found at the base of the Lower Supergroup south of Kirkland Lake (*see* Figure 5, Table 1; Jensen 1978c). Prior to the recognition of komatiitic lavas in the Lower Supergroup, these rocks were designated as the Catherine Basalts (Ridler 1970). This komatiitic succession conformably overlies the Pacaud Tuffs and appears to represent the beginning of a volcanic cycle which culminated in the development of the Skead Group. A tholeiitic succession (called the Catherine Group) separates the Wabewawa Group from the Skead Group.

Preliminary mapping of the Wabewawa Group shows that the group contains olivine spinifex textured peridotitic komatiite lava flows as well as skeletal pyroxene-textured basaltic komatiite flows interlayered with some tholeiitic basalt flows.

Unlike the underlying Pacaud Tuffs, the Wabewawa Group shows little or no tectonic deformation even within 500 m of the Round Lake Batholith (*see* Figure 5). However, dikes of trondhjemite extend from the batholith into the Wabewawa Group.

Similar to the komatiitic succession at the base of the Upper Supergroup, the Wabewawa Group probably represents the encroachment of mafic and ultramafic lava flows on the lower parts of an earlier formed felsic volcanic pile.

GEOLOGICAL INTERPRETATION OF THE KOMATIITIC SUCCESSIONS WITHIN THE LARGER REGION

The distribution of the komatiitic succession at the base of the Upper Supergroup (Stoughton - Roquemaure Group, Lower Formation of the Tisdale Group and Larder Lake Group) (*see* Figures 5 and 49), indicates that komatiitic volcanism occurred over an area of about 10 000 km^2 in the Timmins - Kirkland Lake area. The presence of spinifex textures in the komatiitic lavas suggests that they were derived from local sources (Arndt *et al.* 1977) which may have been the peridotitic intrusions cutting komatiites and immediately underlying felsic volcanic rocks. This distribution implies that komatiitic volcanism occurred throughout the basinal area among the volcanic piles formed earlier (*see* Figure 29). Even though peridotites cut felsic volcanic rocks which were deposited at the edges of the earlier volcanic piles, peridotite does not extend into the underlying rocks which suggests that komatiitic volcanism was restricted to areas where volcanic piles had not developed. The reason may be that peridotitic magmas have greater densities than calc-alkalic rock (Fyfe 1978b).

In the Timmins - Kirkland Lake area, komatiitic volcanism occurred near the time of completion of the earlier volcanic piles. Only in the Larder Lake Group, is there a considerable overlapping of rhyolitic and komatiitic volcanism. The overlapping of komatiitic and calc-alkalic volcanism, however, does not seem to have extended to overlap the extrusion of calc-alkalic basalts and andesites. This suggests that there were time breaks between the cycles of volcanism.

A detailed study of the Stoughton - Roquemaure Group indicates that thicker accumulations of komatiitic and tholeiitic lavas were formed in the basinal area south of the Hunter Mine Group. In this basin, the komatiites probably interfingered with (Pyke 1980), and ultimately covered the sediments derived from the surrounding volcanic piles. As the lavas filled the basin, the floor of the basin started to sink. However, komatiitic volcanism was sufficiently strong at the edges of the volcanic piles that the lavas were partly able to engulf the earlier formed volcanic piles.

While the basin was sinking under the accumulating weight of the komatiitic lavas, much of the differential movement appears to have been concentrated along the edges of the earlier formed volcanic piles in the vicinity of the Destor - Porcupine and Kirkland Lake - Larder Lake Fault Zones. However, no sign of a north-south fault zone having been formed south of Timmins exists to mark the edges of the volcanic pile represented by the Deloro Group. In this area, either the differential downward drop of the basin was more diffuse, or the movement was largely restricted to tilting of the strata to face the centre of the basin.

It is difficult to estimate the thickness of the komatiitic and tholeiitic lavas that were accumulated in the central parts of the basin between the Hunter Mine Group and the Skead Group. If a 10 km thick succession of komatiitic and tholeiitic lavas formed in the type section of the Stoughton - Roquemaure Group which is located on the edge of the older volcanic pile represented by the Hunter Mine Group, then a 30 to 40 km thick sequence may have formed in the basin.

THOLEIITIC VOLCANIC SUCCESSIONS

The base of each tholeiitic succession, represents the cessation of komatiitic volcanism and the beginning of a period dominated by tholeiitic volcanism. During this period, the lavas underwent a strong enrichment in iron. The Kinojevis Group of the Upper Subgroup (*see* Figure 5), which is the main tholeiitic succession of the area, is divided by the Destor - Porcupine Fault Zone into northern and southern sections (Figure 51). The southern section, which includes Goodwin's (1977a, 1977b) Garrison Subgroup (Blake River Group), extends westward into the Timmins area where it has been named the Middle Formation of the Tisdale Group by Pyke (1980; *see* Figure 5). From here, the tholeiitic succession extends southward, around the nose of the synclinorium, and then eastward along the northern side of the Kirkland Lake - Larder Lake Fault Zone toward the Quebec boundary where the rocks were called the McVittie Basalts by Ridler (1970).

The Catherine Group (Jensen 1978c) which Ridler (1970) named the Catherine Basalts, is an older tholeiitic volcanic succession that occurs south of the Kirkland Lake - Larder Lake Fault Zone, in the Lower Supergroup, stratigraphically below the Skead Group (*see* Figure 5).

KINOJEVIS GROUP

Northern Section

North of the Destor - Porcupine Fault Zone, the Kinojevis Group extends west and northwest, from the Quebec boundary into Stoughton and Frecheville Townships where it occupies the core of a triangularly shaped basin (Figure 52). It consists mainly of map units 50 to 500 m thick composed of magnesium-rich and iron-rich tholeiitic basalt that total 4000 m in thickness (Jensen 1978b). The lower 1000 m, however, also contains interflow units of tuff-breccia, crystal tuff, and cherty tuff of calc-alkalic dacite and rhyolite composition, as well as chert, argillite, graphite and cherty oxide iron formation (Figure 53). A total of 78 specimens were analyzed for their major elements; 17 are calc-alkalic interflow units from the lower 1000 m of the section. The Ol'-Ne'-Q'and An-Ab'-Or projections (Figures 54 and 55) of the 78 analyses show that rocks are subalkalic and are orthoclase poor. On the Jensen Cation Diagram (Figure 56), the analyses of the tholeiitic basalts lie in the magnesium- and iron-rich tholeiitic basalt fields and the sedimentary rocks in the calc-alkalic basalt and andesite fields; the tuff and tuff-breccia project in the calc-alkalic dacite and rhyolite fields.

The reason why the thick beds of interflow tuff and interlayered beds of carbonaceous and argillaceous sediment of the lower 1000 m in the Kinojevis Group have the composition of calc-alkalic rocks is not known. Their presence suggests a pause during mafic volcanism, and they may be analogous to the "Middle Marker Horizon" above komatiitic successions in other greenstone belts in Africa and Australia (Viljoen and Viljoen 1969e; Bliss and Stidolph 1969; Durney 1972; Anhaeusser 1973; Fitton *et al.* 1975; Hickman and Lipple 1975; Glikson and Lambert 1976). In the Barberton Greenstone Belt, the Middle Marker was considered by Viljoen and Viljoen (1969g) to indicate a change from komatiitic to tholeiitic volcanism. Later, Williams and Furnell (1979) pointed out that komatiitic lavas are abundant above the Middle Marker in this area. Glikson (1976, 1976c), noting that the Middle Marker was common to many greenstone belts, concluded that it represents the boundary between the Archean oceanic crust and the upper greenstones; he further suggested that there must be an unconformity present at the base or top.

In the upper part of the northern section of the Kinojevis Group, the only interflow material is hyaloclastite, 4 to 40 cm thick. The proportion of iron-rich tholeiitic basalt increases toward the top of the section, indicating an enrichment in iron in the 4000 m section (Map P.2433, back pocket). Tholeiitic andesite, dacite, and rhyolite were not detected in the

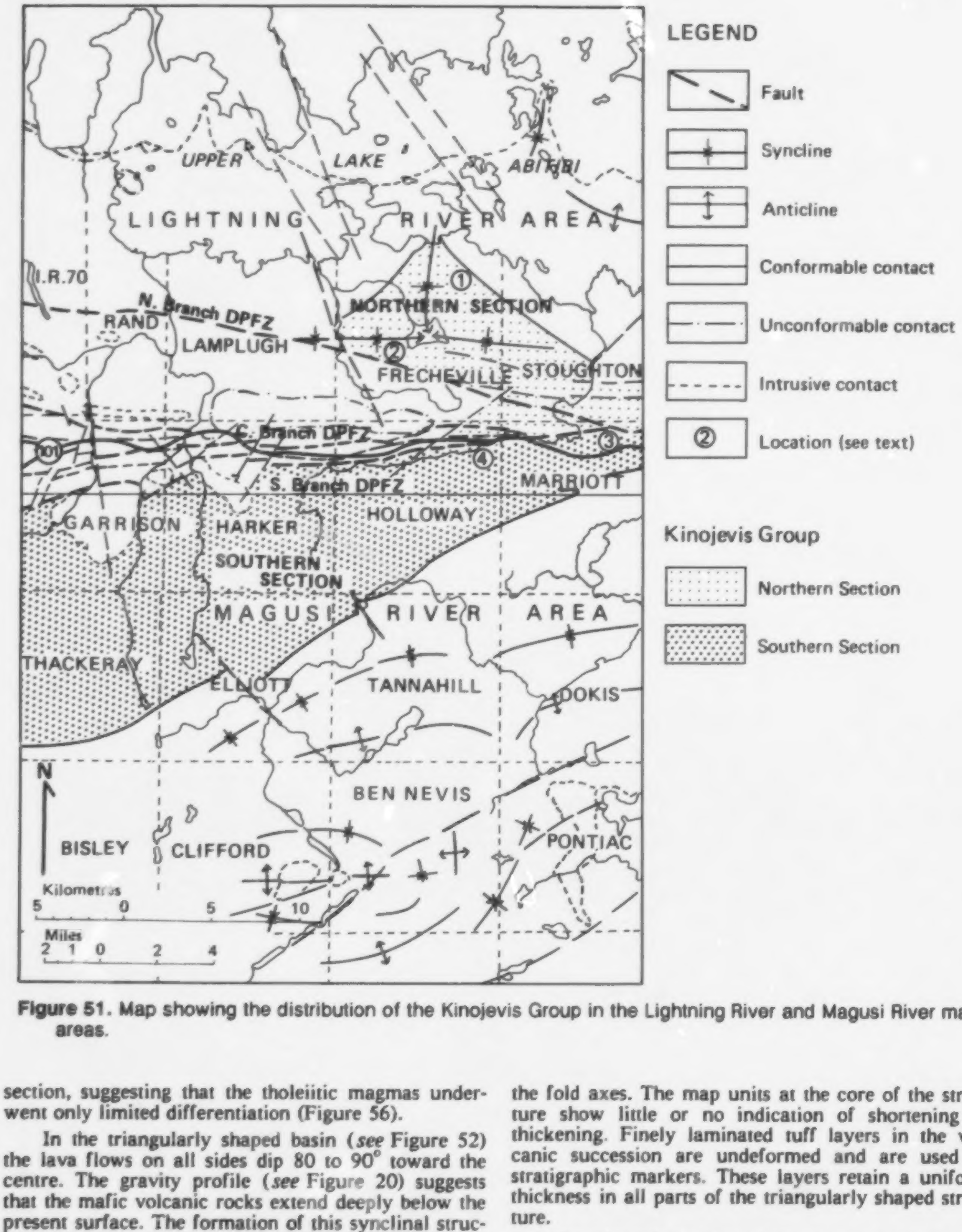


Figure 51. Map showing the distribution of the Kinojevis Group in the Lightning River and Magusi River map areas.

section, suggesting that the tholeiitic magmas underwent only limited differentiation (Figure 56).

In the triangularly shaped basin (*see* Figure 52) the lava flows on all sides dip 80 to 90° toward the centre. The gravity profile (*see* Figure 20) suggests that the mafic volcanic rocks extend deeply below the present surface. The formation of this synclinal structure has been accomplished without penetrative deformation. The lava flows contain pillows, variolites, amygdules, mineral grains, and hyaloclastite fragments which show little or no evidence of stretching or flattening. The only trace of movement is the presence of chlorite-rich slickensides at the flow contacts, and in some pillow selvages of the lava flows at the fold axes. The map units at the core of the structure show little or no indication of shortening or thickening. Finely laminated tuff layers in the volcanic succession are undeformed and are used as stratigraphic markers. These layers retain a uniform thickness in all parts of the triangularly shaped structure.

The triangularly shaped structure must have formed contemporaneously with the volcanism. The very low grade of metamorphism indicates that even the lower strata, including the Stoughton - Roquemaure Group, were never deeply buried by overlying lavas. This indicates that as the volcanic succession developed, each succeeding map unit was contained in a

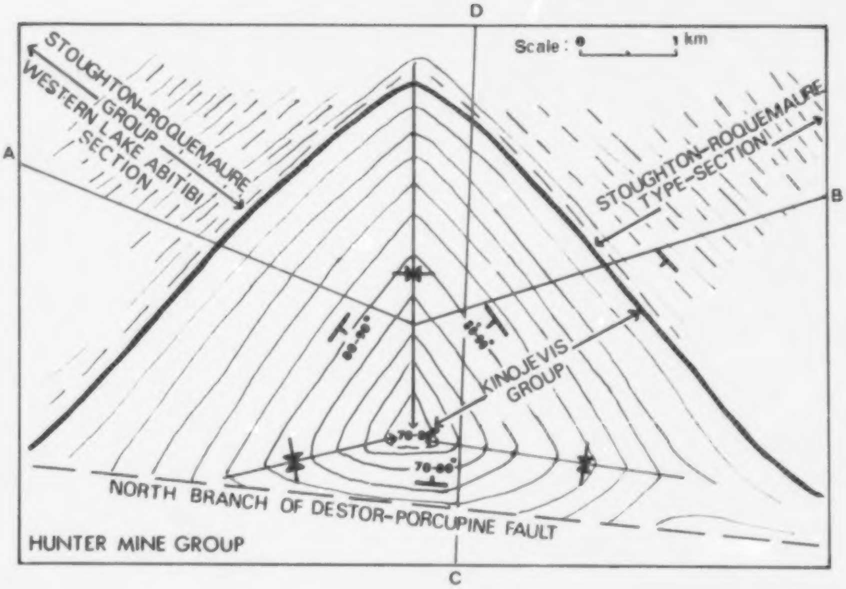

D
Scale : 0 1 km
STOUGHTON-ROQUEMAURE GROUP
WESTERN LAKE ABITIBI SECTION
A
STOUGHTON-ROQUEMAURE TYPE-SECTION
B
KINOJEVIS GROUP
70-80°
70-80°
NORTH BRANCH OF DESTOR-PORCUPINE FAULT
HUNTER MINE GROUP
C

Figure 52. Diagram of the triangularly shaped synclinal fold in Frecheville Township located between location 1 and 2 on Figure 51. Tops face the centre. Cross-sections A-B and C-D are illustrated in Figure 57.

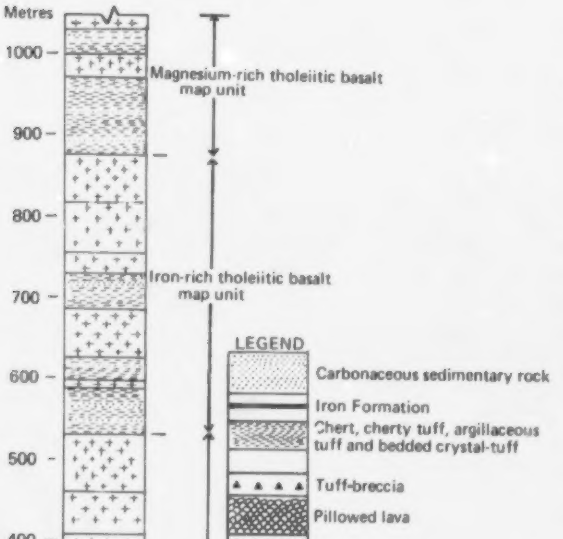

Metres
1000
900
800
700
600
500
400
300
200
100
0
Magnesium-rich tholeiitic basalt map unit
Iron-rich tholeiitic basalt map unit
Magnesium-rich tholeiitic basalt map unit
Iron-rich tholeiitic basalt map unit
Base of Kinojevis Group
Iron-rich tholeiitic basalt
LEGEND
Carbonaceous sedimentary rock
Iron Formation
Chert, cherty tuff, argillaceous tuff and bedded crystal-tuff
Tuff-breccia
Pillowed lava
Massive lava

Figure 53. Stratigraphic column of the lower 1000 m of the Kinojevis Group.

progressively smaller area toward the core of the triangular structure.

Few if any mafic and ultramafic intrusions which could be considered as feeders for the lava flows were found in the volcanic successions making up the sides of the triangular structure. The only intrusions of peridotite and gabbro in the area are those at the base of the Stoughton - Roquemaure Group on the northeastern side. Therefore, the feeders for the younger lavas in the centre probably developed toward the centre of the triangular structure, or else they should have been recognized as intrusions cutting across the older lavas. Since the outer map units do not show stretching and segmentation outward from the centre, the structure could not have been a "spreading center".

The origin of the triangular synclinal structure through the processes of volcanism, is thought to have occurred as shown in Figure 57a. In the first stage, komatiitic lavas began to fill a low area or a shallow basin located in this case, on the edge of the felsic volcanic pile represented by the Hunter Mine Group. During this stage, the sources of lava were dispersed throughout the area of the basin and the lava sheets thickened toward the centre of the basin.

In the second stage (Figure 57b), the floor of the basin sinks causing the lava sheets to pull apart and segment at depth near their point of maximum subsidence. Simultaneously, a magma chamber begins to form in the void caused by the sinking lavas and volcanism concentrates toward the centre of the basin.

At the third stage (Figure 57c), the magma fractionates at depths of about 15 km in a manner similar to the fractionation seen in ocean-ridge spreading centres (Figure 58). A sheeted dike complex may even be present below the central parts of the basin similar to that observed in ophiolite sequences (Coleman 1977). From this point on, komatiitic volcanism does not occur since the peridotitic and picritic magmas frac-

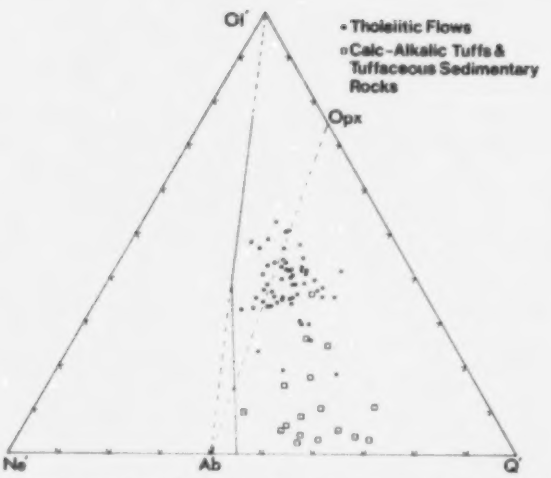


Figure 54. Ol'-Ne'Q' projections of chemical analyses from the northern section of the Kinojevis Group. Solid squares are tholeiitic flows and open squares are calc-alkalic tuffs and tuffaceous sedimentary rocks.

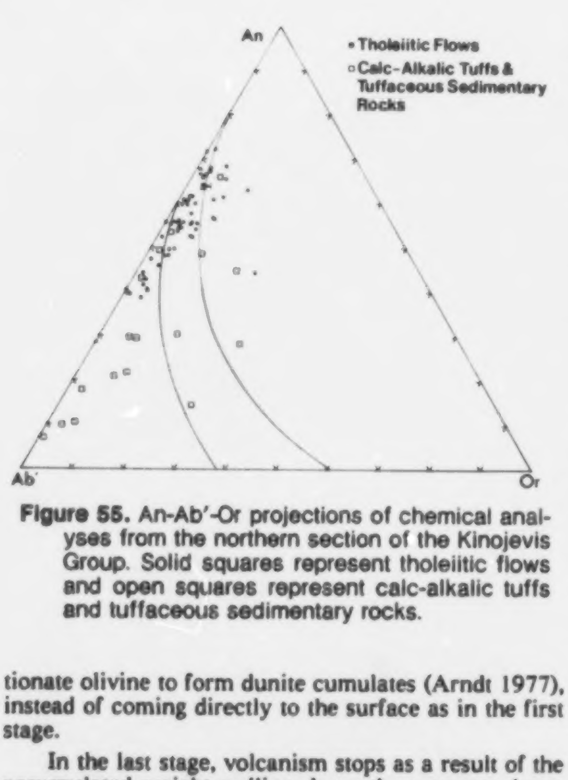


Figure 55. An-Ab'-Or projections of chemical analyses from the northern section of the Kinojevis Group. Solid squares represent tholeiitic flows and open squares represent calc-alkalic tuffs and tuffaceous sedimentary rocks.

tionate olivine to form dunite cumulates (Arndt 1977), instead of coming directly to the surface as in the first stage.

In the last stage, volcanism stops as a result of the accumulated weight pulling down the youngest lavas at the centre of the basin so that these lavas are steeply dipping and arched against each other. This 'closing' of the basin would prevent further sinking and also choke the volcanic vent (Figure 57d) causing the basin to become inactive.

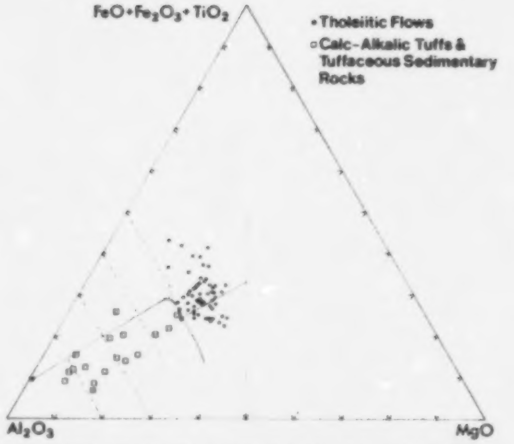


Figure 56. Projection of chemical analyses from the northern section of the Kinojevis Group on the Jensen Cation Diagram. Solid black squares represent tholeiitic flows, and open squares represent calc-alkalic tuffs and tuffaceous sediments.

As the thickest accumulations of lava would be expected to be in the centre of the basin, the measured stratigraphic thickness of 15 km on the northeastern side of the triangular synclinal structure probably represents a minimum thickness of the volcanic deposition. The total cumulated thickness at the centre of the basin may have been 2 or more times greater. However, the arc of tilting (Figures 57a, 57b, 57c, and 57d) suggests that the depth to which the lavas could extend would be about 15 km. Also, the gravity profile indicates that the lavas extend to a depth of about 8 to 10 km. The possibility that the lavas grade into granitic rock at this depth will be discussed at another point further in the report.

Southern Section

The southern section is a steeply dipping south-facing monoclinal succession (Jensen 1978a, 1978b). In Marriott Township, the northern and southern sections of the Kinojevis Group come together and appear to form a continuous succession across the Destor - Porcupine Fault Zone (Figure 51, Location 3). West of Marriott Township, the sections become separated by wedges of older and younger rock preserved between the northern and southern branches of the Destor - Porcupine Fault Zone so that the base of the southern section is truncated (Figure 51, Location 4). Westward from Holloway Township, the exposed stratigraphic thickness increases from about 6000 m to 11 500 m.

In the southern section, 79 analyses plotted on Ol'-Ne'-Q'and An-Ab'-Or diagrams (Figures 59 and 60) show that the rocks are like those to the north. The few analyses that lie in the alkalic and orthoclase fields are of specimens taken near intrusions of quartz monzonite and syenite and are believed to have been enriched in alkalies. The Jensen Cation Diagram shows a tholeiitic suite composed of magnesium-rich and iron-rich tholeiitic basalt, tholeiitic andesite,

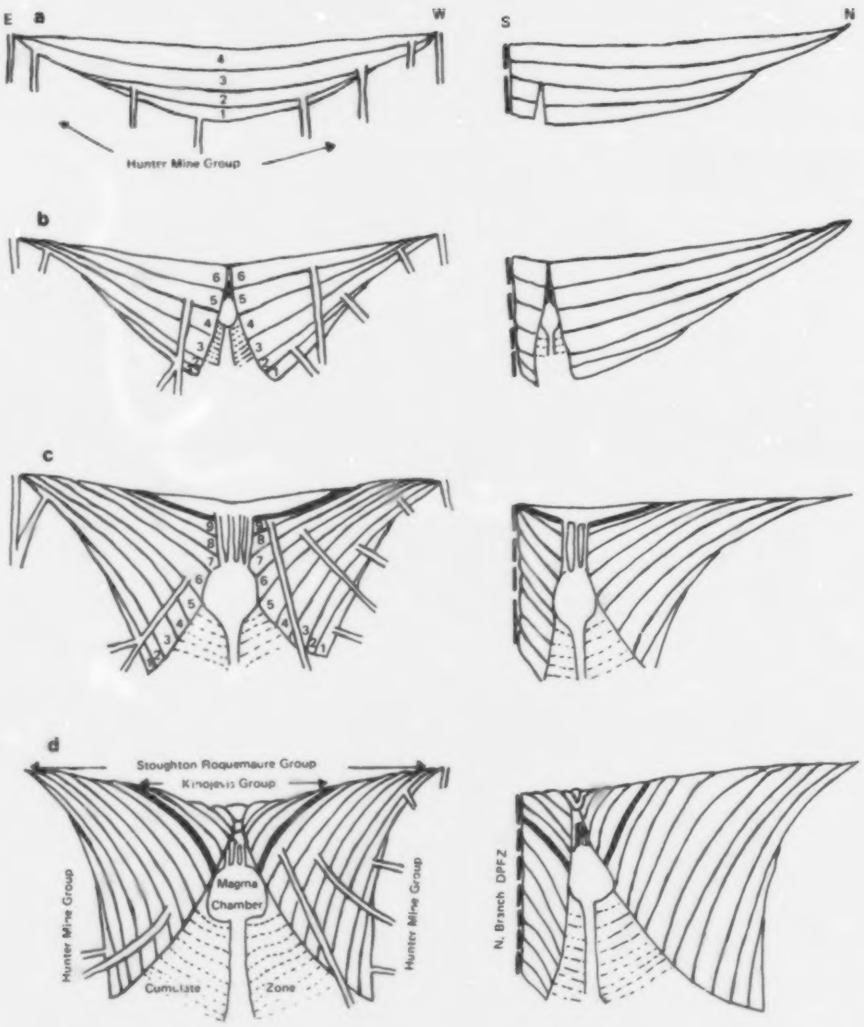


Figure 57. Sketch diagram illustrating the development of the triangularly shaped synclinal fold structure in E-W and N-S directions (profile lines are shown on Figure 52):
a. Komatiitic lava filling the low area on the edge of the felsic volcanic pile represented by the Hunter Mine Group;
b. Transition from komatiitic volcanism to tholeiitic volcanism during deposition of the Stoughton - Roquemaure Group;
c. Initial stage of the volcanic deposition of the Kinojevis Group;
d. Final stages of the volcanic deposition of the Kinojevis Group.

dacite, and rhyolite (Figure 61). The 3 analyses in the calc-alkalic basalt field are from magnesium-rich tholeiitic basalt flows also rich in feldspar phenocrysts and the other 4 in the calc-alkalic field are sedimentary rocks.

The iron content of the tholeiitic basalts increases toward the top of the section. Increasing amounts of tholeiitic andesite, dacite and rhyolite towards the top of the section indicate that differentiation was able to go to completion (Green and Ringwood 1967a), in contrast to the northern section (*see* Figure 56).

Kinojevis Group in the Western End of the Synclinorium

North from the Watabeag Batholith (*see* Figure 5), the base of the Kinojevis Group is exposed south of the Destor - Porcupine Fault Zone where it has been designated as the Middle Formation of the Tisdale Group (Pyke 1978a, 1978b). Calc-alkalic rhyolitic tuff and tuff-breccia, and iron formation, combine to form interflow units with the tholeiitic basalts at the base of the succession. Upward, the calc-alkalic rocks and iron formation change to bedded cherts containing laminae of carbon and clay. About 1000 m stratigraphically above the base, the volcanic succession consists mainly of tholeiitic basalt flows like those found elsewhere in the Kinojevis Group.

The calc-alkalic tuff-breccias and iron formation at the base of the Middle Formation have not been reported farther west by Pyke (1982). Instead, he placed the base of the Middle Formation above the komatiitic lavas and below the appearance of abundant iron-rich tholeiitic basalt flows. Using this criterion, Pyke (1982) was able to follow the contact around the nose of the synclinorium and as far south as the Kirkland Lake - Larder Lake Fault Zone. The top of the Kinojevis Group can also be traced on the regional aeromagnetic maps, Maps 7084G and 7085G (Ontario Department of Mines - Geological Survey of Canada 1970) because the iron-rich tholeiitic basalts have a higher aeromagnetic expression than do the overlying calc-alkalic rocks.

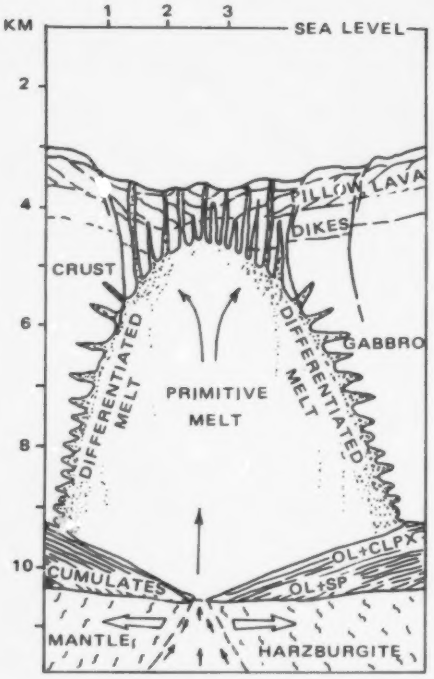


Figure 58. Model of inferred magma beneath oceanic spreading ridge (after Bryan and Moore 1977).

Kinojevis Group on the Southern Limb of the Synclinorium

On the southern limb of the synclinorium, the Kinojevis Group strikes southeastward into the Kirkland Lake - Larder Lake Fault Zone, where it is unconformably overlain by rocks of the Timiskaming Group (*see* Figure 5). Since the Kinojevis Group does not extend south of the Timiskaming Group, it is considered to be truncated by the fault beneath the Timiskaming Group. In this area, the base of the Kinojevis Group is considered to have been downfaulted along the northern side of the Kirkland Lake - Larder Lake Fault Zone.

DISTRIBUTION OF THE KINOJEVIS GROUP BETWEEN THE DESTOR - PORCUPINE AND KIRKLAND LAKE - LARDER LAKE FAULT ZONES

The eastward plunging syncline of the Blake River Group, Kinojevis Group (Middle Formation of the Tisdale Group) and the Lower Formation of the Tisdale Group can be observed in Figure 62. The change from komatiitic to tholeiitic volcanism probably came about when the lower komatiitic and tholeiitic lavas sank, as shown in Figure 57, thus allowing the formation of magma chambers in which olivine and pyroxene fractionated and remained at depth instead of being extruded as part of the komatiitic and picritic lavas. During the tholeiitic volcanism, the western (Timmins) part of the basin became inactive, and the volcanic activity concentrated in the Kirkland Lake area until the start of calc-alkalic volcanism.

The rate of thickening of the lava sheets toward the centre of the basin is unknown. If the lavas thickened inward 2 to 3 times, a tholeiitic succession 20 to 30 km thick may have been deposited above an equally thick komatiitic succession between the 2 fault zones.

CATHERINE GROUP

South of the Kirkland Lake - Larder Lake Fault Zone, the Catherine Group was probably formed in a basin filled with komatiitic lavas of the Wabewawa Group. To accommodate the Catherine Group and to cause the transition from komatiitic to tholeiitic volcanism, the floor of the basin must have been sinking under the weight of the accumulated komatiitic lavas in a manner similar to that shown in Figures 57 and 63.

According to Ridler (1969), the Catherine Group thickened eastward by the addition of imbricated lava flows. However, the Catherine Group has not been mapped in detail to determine the exact geometry of the lava flows. Also, the rocks east of the Lower Supergroup are obscured by Proterozoic rocks and the Pontiac Massif occurs east of them (*see* Figure 4). A northeast-trending fault zone below the Proterozoic rocks may cut the lavas off as shown in Figure 63.

GEOLOGICAL ANALYSES OF THE THOLEIITIC SUCCESSIONS

The tholeiitic successions were preceded by komatiitic successions composed of lavas derived from widely scattered feeders. When the komatiitic successions became thick enough, they subsided and the sources of volcanism migrated toward the point of maximum subsidence. The change from komatiitic to tholeiitic volcanism was probably related to the development of a central magma chamber at a depth of 15 to 20 km below the surface in which fractionation occurred. Once this happened, the magmas were funnelled through it and komatiitic magmas could no longer be extruded (Figure 57).

In the Kinojevis Group, TiO_2, REE, and iron contents of the basalt increase together (Jackson and Smith 1978). In oceanic basalts, Nisbet and Pearce (1977) suggested that these elements increased according to the increase in rate of spreading at oceanic ridges because this increases the ratio of the volume of the chamber to the volume of an individual flow. In the Kinojevis Group, where spreading was not evident, the increase in TiO_2, and REE, and the formation of the iron-rich tholeiites was probably related, instead, to the progress of sinking of the basin. Stepwise, a faster rate of sinking and spreading of the lava sheets at depth would have the same effect (Figure 57).

Lavas of the tholeiitic successions and those in the upper parts of the underlying komatiitic successions probably were deposited in shallow seas at levels equal to the shelves of the earlier formed calc-alkalic volcanic piles (*see* Figure 27). The general lack of interflow sedimentary rock in the tholeiitic successions supports the concept that the lavas were not deposited in deep basins adjoining areas of high relief. The lavas were not deposited around shield-type volcanoes which would have caused the lavas to spread out and cover wider areas as volcanism progressed.

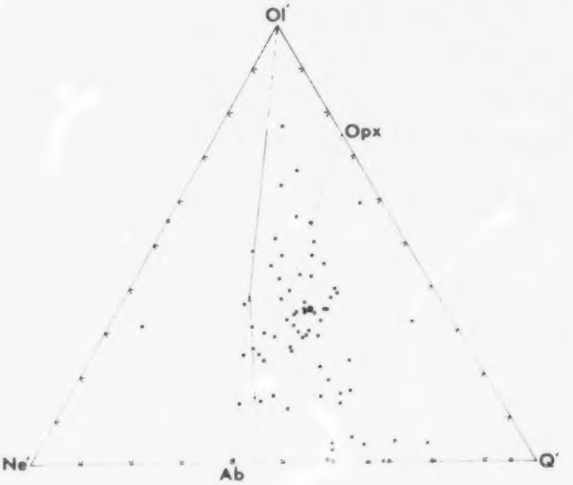


Figure 59. Ol'Ne'-Q' projections of chemical analyses from the Kinojevis Group south of the Destor - Porcupine Fault.

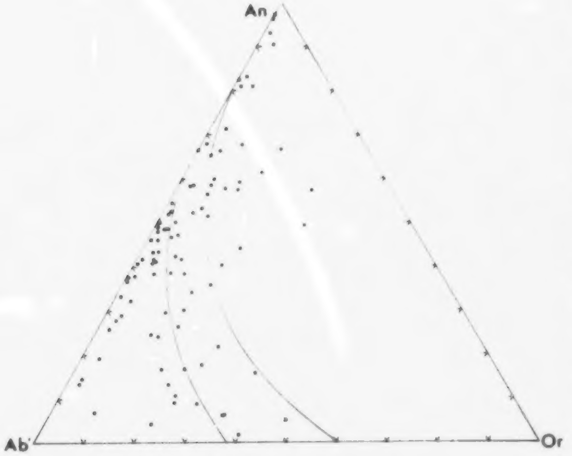


Figure 60. An-Ab'-Or projections of chemical analyses from the southern section of the Kinojevis Group.

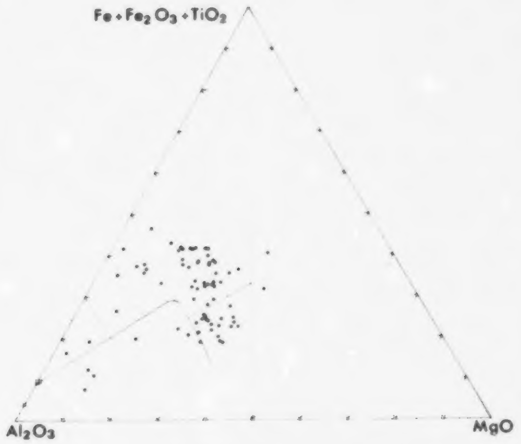


Figure 61. Projections of chemical analyses from the southern section of the Kinojevis Group on the Jensen Cation Diagram.

Instead, the extent of the succeeding lava sheets contracted toward the centre of volcanism, and these lavas gradually "closed" their basin to further volcanism. The basin of volcanism north of the Destor - Porcupine Fault Zone was small enough that it quite quickly became inactive before felsic tholeiitic and calc-alkalic volcanism developed, as compared to the basin south of the Destor - Porcupine Fault Zone which allowed subsequent calc-alkalic volcanism. The latter basin was much larger, and therefore appears to have been able to accommodate a much thicker succession of tholeiitic and komatiitic lava flows than the basin north of the Destor - Porcupine Fault Zone. A similar observation may also be applied to the basin in which the Catherine Group was deposited, as these rocks are overlain by calc-alkalic volcanic rocks of the Skead Group.

THE CALC–ALKALIC VOLCANIC SUCCESSION OF THE UPPER SUPERGROUP (BLAKE RIVER GROUP)

The Blake River Group which overlies the Kinojevis Group in the core of the synclinorium (*see* Figure 5) represents the period of calc-alkalic volcanism following tholeiitic volcanism. The Blake River Group is a calc-alkalic volcanic succession composed mainly of basalt and andesite, with dacite and rhyolite. It extends westward 140 km from its original type section in Cadillac Township, Quebec, into Ontario as far as Benoit Township (Figure 64). On the Ontario side of the provincial boundary, the group is about 10 km thick and occupies an area 32 km wide. In Quebec, where the larger part of the Blake River Group occurs, almost all of the rocks between the Destor - Porcupine Fault Zone and the Kirkland Lake - Larder Lake Fault Zone are included in the group.

In other areas of the world, Barker and Peterman (1974), Shaw (1976), and Glikson (1976a) indicate that, in general, there is a paucity of calc-alkalic basalts and andesites in Archean greenstone belts, particularly of thick successions of these rocks such as the Blake River Group. The Blake River Group represents a volcanic succession that was either rarely formed, or was rarely preserved in other greenstone belts.

At this point, it should also be noted that the thick tholeiitic successions, like the Kinojevis Group, containing highly fractionated iron-rich tholeiitic basalt as well as tholeiitic andesite, dacite, and rhyolite are also rarely found in other greenstone belts (Gill and Bridgewater 1979). Therefore, it would seem that the presence of thick, highly differentiated tholeiitic volcanic successions may be closely related to the presence of thick calc-alkalic volcanic successions.

The Blake River Group in Ontario (Magusi River area) is conveniently divided into a layered assem-

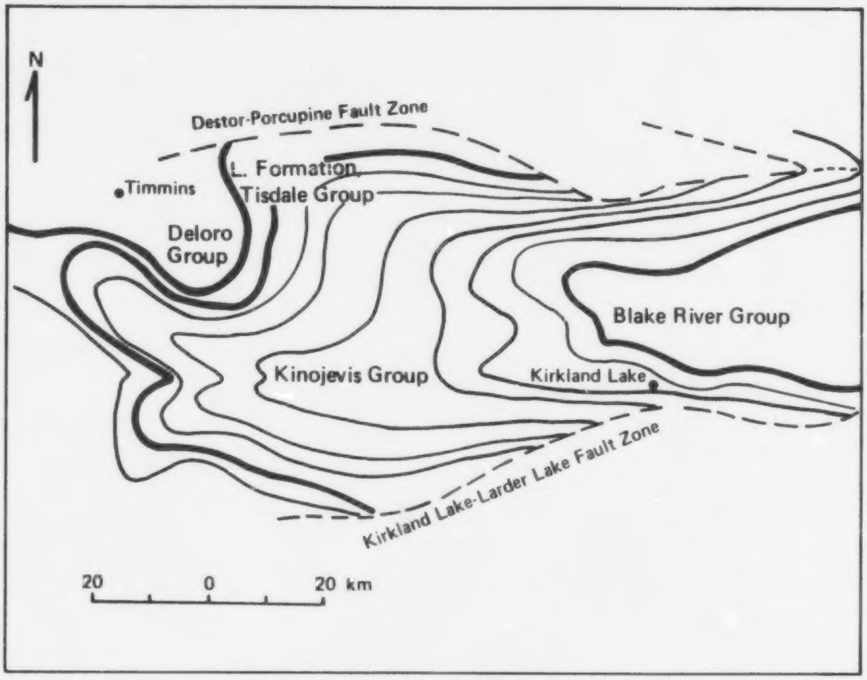


Figure 62. Diagram of the distribution of the Kinojevis Group in the basin of volcanism. Note the concentric distribution of the Lower Formation of the Tisdale Group and Kinojevis - Blake River Groups in the synclinorium. Lines in the Kinojevis Group are depositional isochrons.

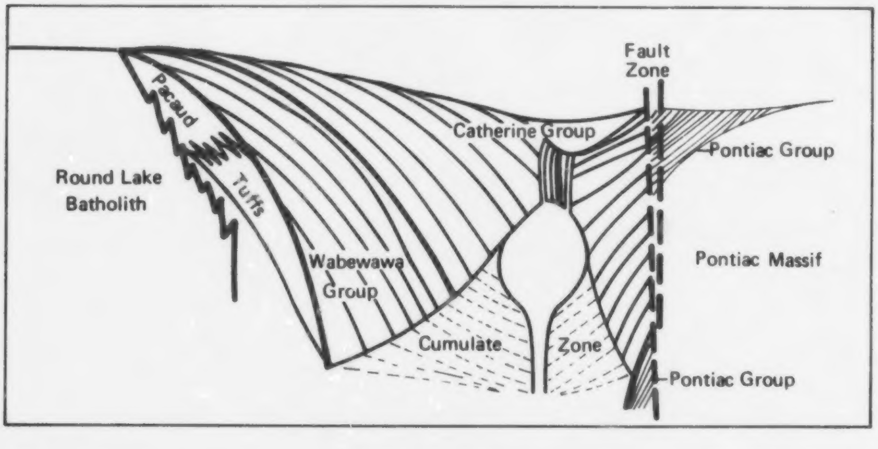


Figure 63. Diagram illustrating the deposition of the Catherine Group in the Lower Supergroup.

blage, an intrusive assemblage, and related structural features. Each of these aspects of the Blake River Group differs considerably from those found in the groups discussed previously, including the calc-alkalic successions found in the lower supergroups.

Several detailed studies have been made on the Blake River Group in the Noranda Area, Quebec (Dimroth 1975a; Dimroth and Demarke 1978; Dimroth and Rocheleau 1979; Larson and Webber 1977; Goodwin 1977, 1979; Goldie 1979a, 1979b; Rosen-Spence 1976; Gelinas *et al.* 1977a). Only a few of the large-scale features of the Noranda part of the Blake River Group pertaining to the development of the Kirkland Lake part of the Abitibi Belt will be discussed in this report.

BLAKE RIVER GROUP IN ONTARIO (MAGUSI RIVER AREA)

Layered Assemblage

In the Magusi River map area, the layered rocks of the Blake River Group are composed of lava flows, pyroclastic rocks, and sedimentary rock formed of volcanic detritus that form a 10 km thick volcanic succession which has been described by Jensen (1972, 1975a, 1975b, 1978a, 1978b). The rocks range from magnesium-rich tholeiitic basalt to calc-alkalic rhyolite, with calc-alkalic basalt and andesite being dominant. The layered rocks are intruded by numerous stocks of gabbro, diorite, and quartz diorite as well as rhyolite domes, and are folded (Jensen 1975a, 1978a).

Whole-rock major-element analyses were made of 208 specimens from the lava flows, pyroclastic rocks, sedimentary rocks, rhyolite domes, and coarse-grained intrusive rocks. Of these, about 84 are from the lower half of the Blake River Group and 87 are from the area approximately covering the upper half of the group (Figure 65). The Ne′-Ol′-Q′projections of the analyses from the approximate lower and upper halves of the Blake River Group (Figures 66 and 67, respectively) show that quartz normative subalkalic volcanic rocks dominate in the Blake River Group. The An′-Ab′-Or projections (Figures 68 and 69, respectively) show that the rocks tend to be orthoclase-poor, and Figure 69 indicates a weak orthoclase enrichment of the rocks upward in the Blake River Group.

On the Jensen Cation Diagram (Figures 70 and 71), the analyses project as a calc-alkalic volcanic suite composed mainly of calc-alkalic basalts and andesites with some calc-alkalic dacite and rhyolite and some magnesium-rich tholeiitic basalt. In the upper half (Figure 71), the projections form 2 distinct populations, 1 centred in the calc-alkalic basalt and andesite fields and the other, in the calc-alkalic rhyolite and dacite fields. Two similar, less distinct, populations also occur on Figure 70 suggesting that there is a hiatus in the rock compositions between the felsic calc-alkalic volcanic rocks and the intermediate to mafic calc-alkalic rocks.

In Marriott Township (*see* Figure 65, Location 1), the base of the Blake River Group is composed mainly of pillowed variolitic basalt flows ranging from magnesium-rich tholeiitic basalt to calc-alkalic basalt. Dimroth *et al.* (1974) extends this unit for several km eastward into Quebec along the base of the Blake River Group. West of Marriott Township, this variolitic unit wedges out where younger calc-alkalic basalt and andesites of the Blake River Group overlie iron-rich tholeiitic basalt of the Kinojevis Group. This suggests that the calc-alkalic volcanism may have been centred in Quebec and spread eastward into Ontario. In places, the calc-alkalic basalts appear to interfinger with the iron-rich tholeiitic basalts and minor discontinuous units of iron-rich tholeiitic basalt pillowed lava are interlayered with the calc-alkalic basaltic and andesitic lavas within the lower 1000 m of the Blake River Group (*see* Figure 65, Locations 2, 3, 4, and 5) suggesting gradual change from tholeiitic to calcalkalic volcanism. The iron-rich tholeiitic basalt flows, interlayered with the calc-alkalic lavas, do not form thick tabular flows nor are they as iron-rich as those in the underlying Kinojevis Group.

The lower half of the Blake River Group consists mainly of calc-alkalic basalt and andesite, and magnesium-rich tholeiitic basalt in the form of tabular lava flows, pillowed lavas, and pillow breccias that form map units 30 to 500 m thick and 0.6 to 8.0 km long. Calc-alkalic dacite and rhyolite is limited to lenses less than 1000 m long and 300 m thick (*see* Figure 65, Locations 6, 7, 8).

The upper half of the Blake River succession appears to have been deposited around volcanic domes containing cores of massive rhyolite cut by quartz-diorite in Clifford and Pontiac Townships (Jensen 1975a, 1975b), (*see* Figure 65, Locations 9 and 10). Smaller satellite bodies of massive dacite and rhyolite occur in Ben Nevis and Dokis Townships (Figure 65, Locations 11 and 12). In the cores of the volcanic dome, the rhyolite is mainly a massive, fine-grained, siliceous rock, which contains quartz and plagioclase (An_{20} to An_{40}) phenocrysts, 1 to 3 mm in size (Jensen 1975a, 1975b). The plagioclase grains are euhedral and zoned. The quartz grains are bipyramidal-shaped and have deep embayments of matrix material indicating that the quartz grains have undergone partial resorption during their transport to the surface (Green and Ringwood 1967a, 1967b).

Adjacent to the volcanic domes, the rhyolite consisted of flow-breccia and tuff-breccia. Away from the domes, the rhyolite tuff-breccias grade into fine-grained bedded crystal tuffs and cherty tuffs (Jensen 1975a, 1975b), suggesting that the volcanic domes were the sources for the layered felsic volcanic rocks.

Tabular lava flows, pillowed lava and pillow-breccia of magnesium-rich tholeiitic basalt, calc-alkalic basalt and andesite occur interlayered with the rhyolite and dacite tuff and tuff-breccia. Many of these lavas are much more amygdaloidal than the lavas in the lower half of the Blake River Group and indicate that they were formed in shallower water (Sigvaldason 1968; Wells *et al.* 1979). The amygdules are up to 3 cm in size, and filled with quartz, chlorite, prehnite, pumpellyite, and albite, and form up to 50% of the volume of some rocks.

A 3 km thick and 10 km long, lens-shaped zone of bedded volcanic detritus occurs north of the volcanic domes (*see* Figure 65, Location 13). In it, the beds are from 5 cm to 50 m thick and are composed of subrounded to angular clasts of calc-alkalic basalt and andesite, 0.5 to 60 cm in size, set in a finer grained matrix of crystals, fragments of chloritic glass, and lithic rock particles cemented together by chlorite, albite, quartz, and prehnite. The clasts are variably sorted. Where graded bedding is present, it may be inverse or normal and the clasts may be weakly imbricated. Some of the beds have channelling which may have been caused by turbidity currents (Walker 1976) probably caused by volcanic debris from the volcanic domes to the south.

Overall, the layered rocks of the Blake River Group indicate a changing depositional environment from one of low, possibly slightly negative relief represented by the Kinojevis Group to one of positive relief with prominent volcanic edifices. This change may have been caused by the calc-alkalic rocks being

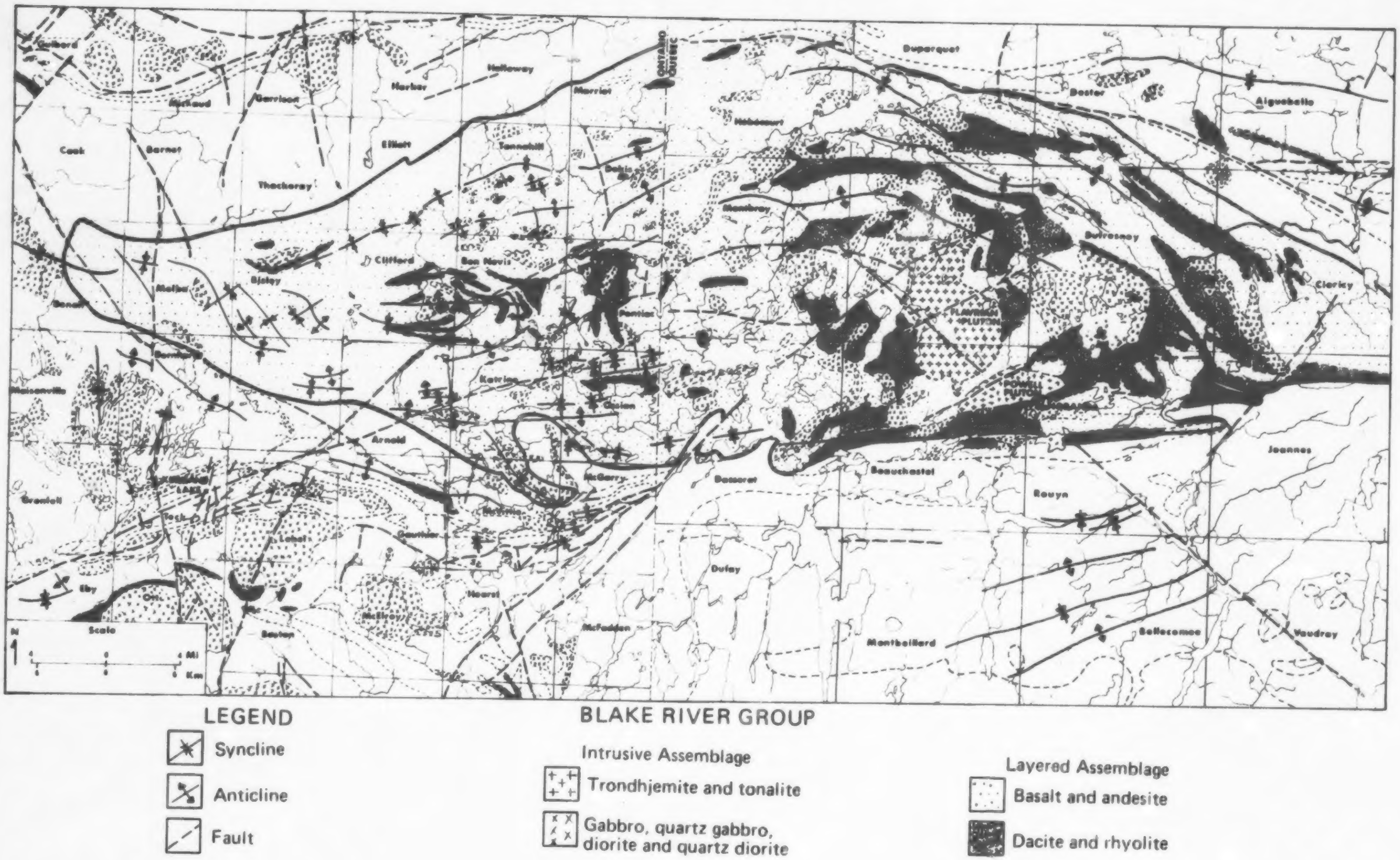


Figure 64. Map of the Blake River Group in Ontario and Quebec (Quebec part of the Blake River Group after Grenier 1967). Note type section of the Blake River Group occurs in Cadillac Township, 20 km east of Joannes Township.

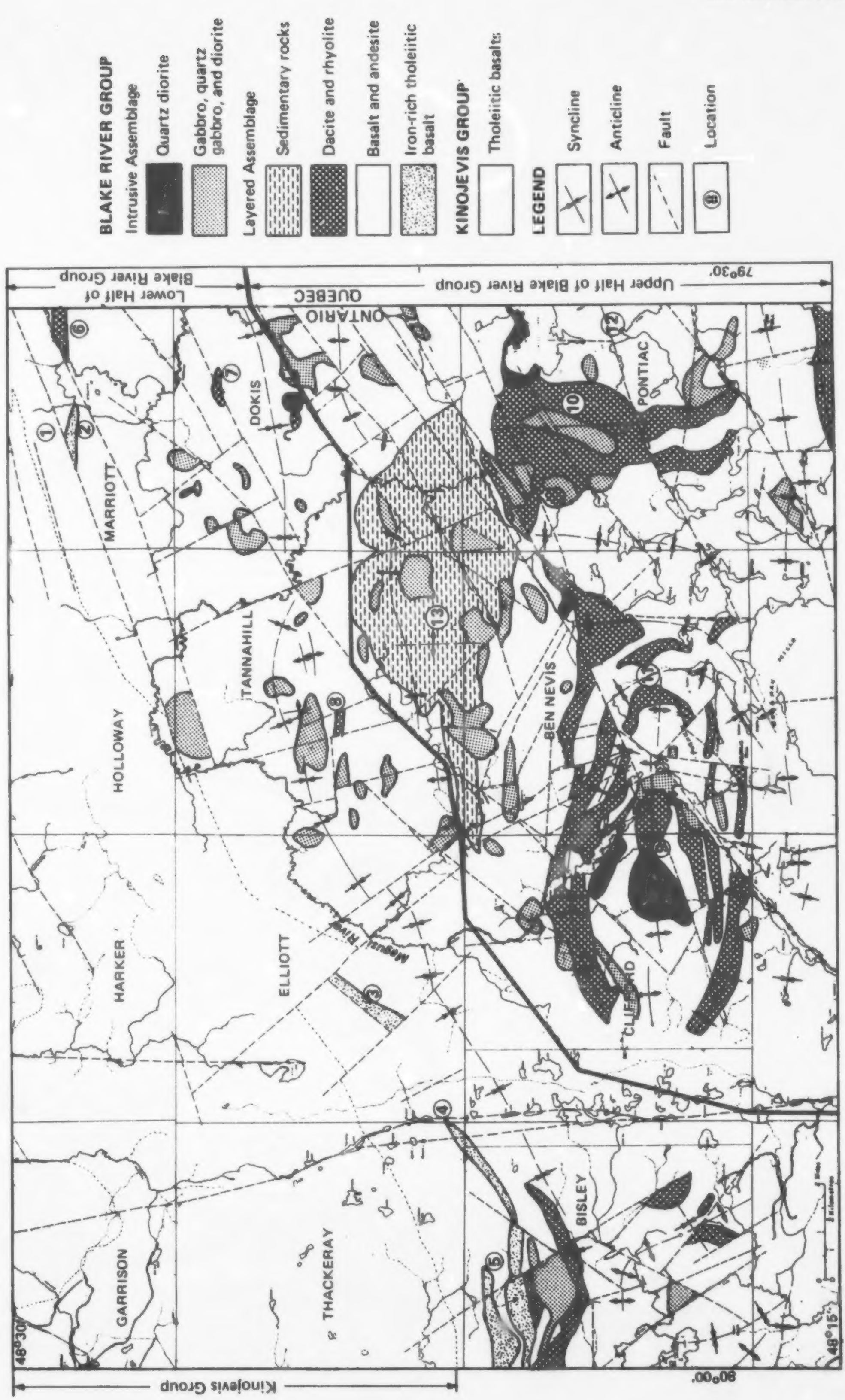


Figure 65. Map of Blake River Group in the Magusi River map area.

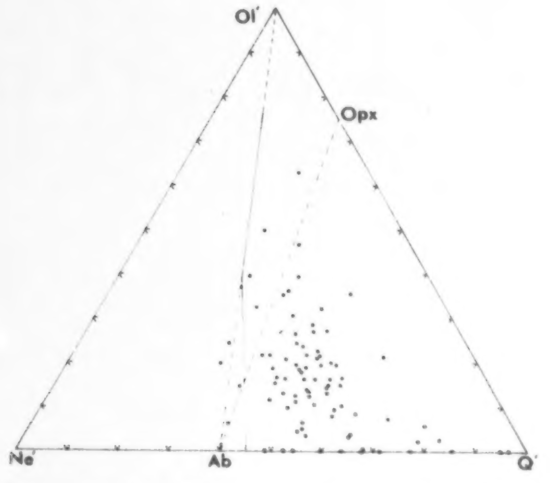


Figure 66. Ne'-Ol'-Q' projections of analyses of the layered rocks from the lower half of the Blake River Group.

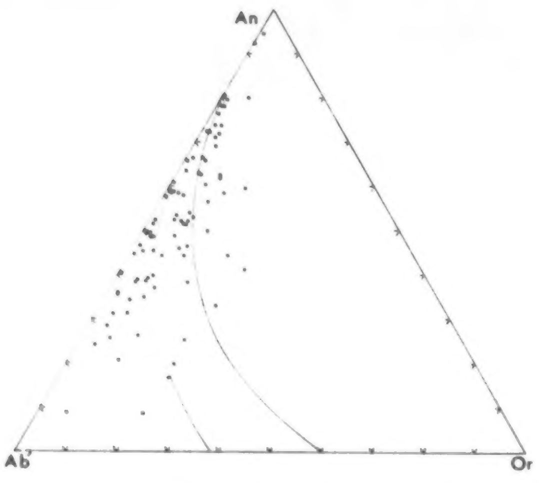


Figure 68. Ab'-An-Or projections of chemical analyses of the layered rocks from the lower half of the Blake River Group.

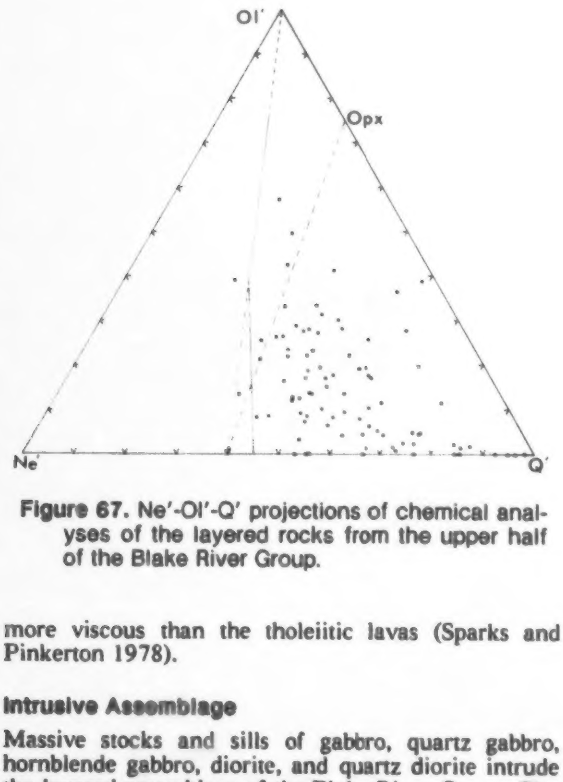


Figure 67. Ne'-Ol'-Q' projections of chemical analyses of the layered rocks from the upper half of the Blake River Group.

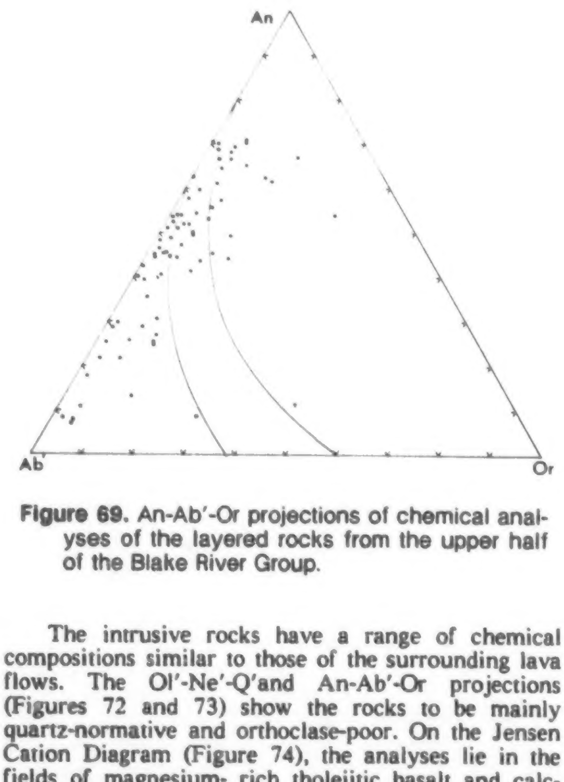


Figure 69. An-Ab'-Or projections of chemical analyses of the layered rocks from the upper half of the Blake River Group.

more viscous than the tholeiitic lavas (Sparks and Pinkerton 1978).

Intrusive Assemblage

Massive stocks and sills of gabbro, quartz gabbro, hornblende gabbro, diorite, and quartz diorite intrude the layered assemblage of the Blake River Group. The stocks, which form the majority of the intrusions, are circular to elliptical in shape and show near vertical intrusive contacts with their layered host rocks which have dips of 50 to 80° to indicate that intrusion occurred after the layered rocks had been folded. The stocks are from 100 to 300 m across and in a few places, they appear to grade into fine-grained tabular lava flows, strongly suggesting the intrusive rocks were feeders for the lava flows.

The intrusive rocks have a range of chemical compositions similar to those of the surrounding lava flows. The Ol'-Ne'-Q'and An-Ab'-Or projections (Figures 72 and 73) show the rocks to be mainly quartz-normative and orthoclase-poor. On the Jensen Cation Diagram (Figure 74), the analyses lie in the fields of magnesium- rich tholeiitic basalt and calc-alkalic basalt and andesite suggesting that these intrusions were part of the calc-alkalic volcanic activity.

In the area underlain by rocks of the lower half of the Blake River Group (Figure 65), the intrusions are mainly gabbros and quartz gabbros whereas in the area underlain by the upper half of the group, diorites are more common. Quartz diorites intrude the rhyolitic cores of the volcanic domes. This corresponds to the field observation that felsic lavas be-

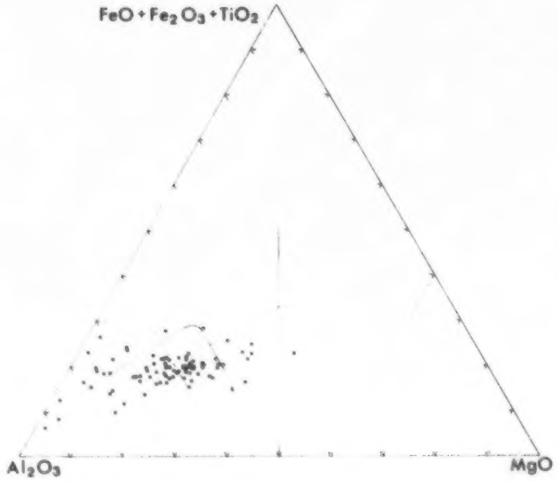


Figure 70. Projections of the chemical analyses of the layered rocks from the lower half of the Blake River Group on the Jensen Cation Diagram.

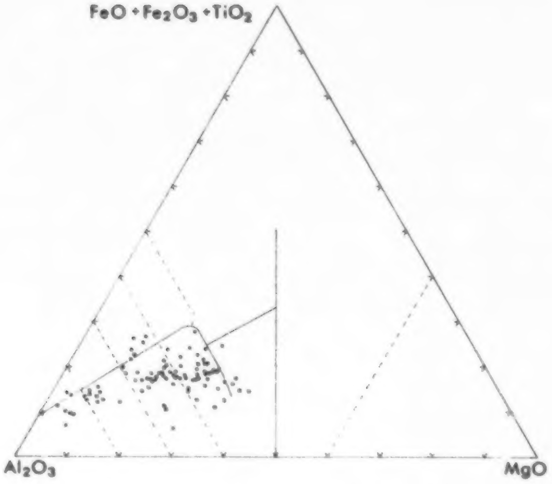


Figure 71. Projections of chemical analyses of the layered rocks from the upper half of the Blake River Group on the Jensen Cation Diagram.

come more abundant upward in the Blake River Group. However, the quartz-diorite stocks which are the most felsic of the intrusive rocks, are equivalent to calc-alkalic andesite (Figure 73) and thus were not feeders for the rhyolites. The source of the rhyolite flows and tuff-breccias was probably the rhyolite bodies in the volcanic domes.

The scattered occurrence of the intrusions suggests that the feeders were not exclusively central vents, but volcanism may have been strongest near the volcanic domes. In the Kinojevis Group and in the Stoughton - Roquemaure Group around Lake Abitibi, observable intrusive rocks that could have acted as feeders for the tholeiitic and komatiitic lavas are scarce. The degree of erosion of the Blake River Group must have been limited because the rocks have prehnite and pumpellyite (Jensen 1975a, 1975b), which indicates that only 2 to 3 km of rock has been removed since their deposition (Jolly 1978). Therefore, the stocks and sills were high-level intrusions, and igneous activity continued throughout the area of lava deposition. The increased stratigraphic thickness towards the centre of the basin is thought to be due to subsidence and greater volcanic activity. The distribution of the stocks suggests that, unlike tholeiitic volcanism which contracted toward a central area, calc-alkalic volcanic activity was likely spreading outward.

Structure

Unlike the lavas of the underlying Kinojevis Group which form a thick monoclinal succession, the layered rocks of the Blake River Group are folded into a series of doubly plunging synclines and anticlines (Jensen 1972, 1975a, 1975b, 1978a). The pillows, amygdules, mineral grains, and chloritic glass fragments, and rock fragments, however, retain their original shapes. The development of cleavages and schistosities in the rocks near the axes of folds is rarely observed, suggesting that penetrative deformation was not important during the folding of the rocks.

The synclines and anticlines produce a concentric fold pattern around the volcanic domes (Figure 65). In Clifford Township (*see* Figure 65), the volcanic rocks dip steeply and face away from the volcanic dome in all directions and form the inner limb of a syncline that encircles the volcanic dome. Similar synclines are found south and west of the volcanic dome in Pontiac Township (*see* Figure 65). Curvilinear anticlines complimentary to the above synclines, occur outward, farther away from the volcanic domes. In contrast, the quartz diorite and massive rhyolite bodies in the volcanic domes are subvertical (Jensen 1975a) suggesting that they remained upright in position during the folding. Many of the gabbroic stocks have subvertical intrusive contacts regardless of the dip of the lavas which they cut to indicate that the stocks intruded the lavas after these had folded. This suggests that igneous activity occurred over a wide area during and after the folding of the lavas.

Petrogenesis

Hawkesworth and O'Nions (1977), O'Nions and Pankhurst (1978), and Ewing (1979) have concluded that Archean calc-alkalic basalts and andesites required for their formation, the partial melting of eclogite or of garnet-bearing amphibolite, both similar in composition to tholeiitic basalt. Ringwood (1975) pointed out that quartz tholeiite and olivine tholeiite would be converted to garnet granulite and eclogite at depths of about 15 km and temperatures of about 300° and that, at 500 to 600°, the conversion would be spontaneous (Figure 75). From this experimental work as well as seismic evidence, Ringwood (1975) concluded that eclogite must be a principal component of rocks in Precambrian Shield areas which occur at depths below 15 km. The Stoughton - Roquemaure Group and the Kinojevis Group between the Destor - Porcupine and Kirkland Lake - Larder Lake Fault

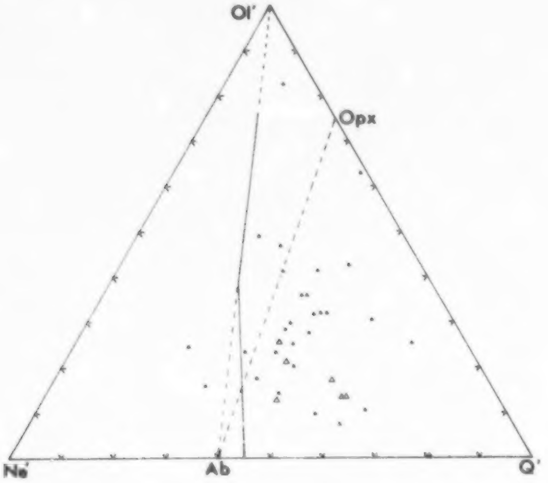


Figure 72. Ne'-Ol'-Q' projections of chemical analyses of the intrusive rocks of the Blake River Group: solid triangles represent gabbro, quartz gabbro, and hornblende gabbro, and open triangles diorite and quartz diorite.

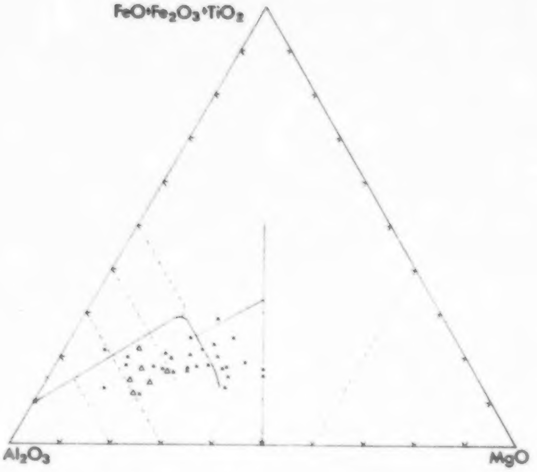


Figure 74. Projections of chemical analyses of the intrusive rocks from the Blake River Group on the Jensen Cation Diagram. Solid triangles represent gabbro, quartz, gabbro, and hornblende gabbro and open triangles, diorite and quartz diorite.

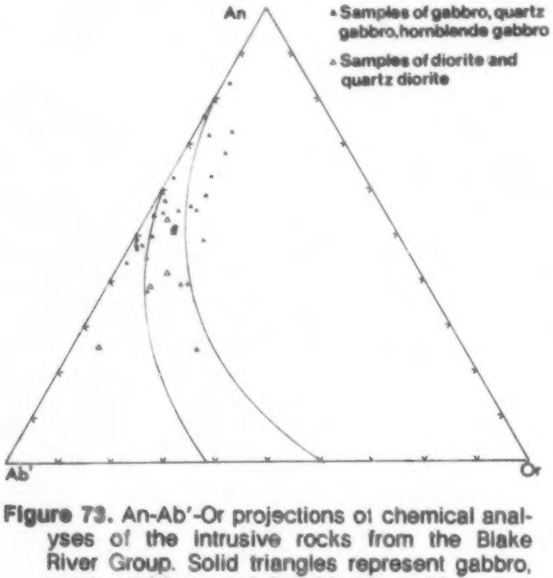


Figure 73. An-Ab'-Or projections of chemical analyses of the intrusive rocks from the Blake River Group. Solid triangles represent gabbro, quartz gabbro, and hornblende gabbro, and open triangles diorite and quartz diorite.

Zones likely formed volcanic successions of 30 to 40 km, and 20 to 30 km thick, respectively. As these thick mafic to ultramafic successions were sinking, their basal parts could have undergone a conversion to garnet-rich amphibolite and ultimately to eclogite. Because of its high density of 3.45, eclogite would have continued to sink farther into the mantle while undergoing partial melting (Green and Ringwood 1967a, 1967b; Figure 76).

The experimental results of Green and Ringwood (1967a, 1967b) to the Blake River Group, indicate that the magmas responsible for these rocks were generated by the partial melting of eclogite of low-H_2O content at depths greater than 100 km. Green and Ringwood (1967a, 1967b) proposed that the very existence of rhyolite with quartz phenocrysts would have required that felsic magmas be produced by the fractionation of garnet and pyroxene from andesitic magmas at depths greater than 100 km so that as the magmas become more siliceous, quartz could replace garnet in the liquidus phase. Under dry conditions, the resultant dominant magma type would be andesitic in composition, because this magma forms at lower temperatures than do more mafic and more siliceous magma compositions, and it is not until the more siliceous phase in the eclogite is exhausted, that more basic magmas can form. If amphibolite-rich rock had undergone partial melting, quartz dioritic and dacitic magmas would have formed at shallow depths resulting in felsic volcanic rocks free of quartz phenocrysts because garnet, and at shallower depths, plagioclase, not quartz, would have been in the liquidus phase (Ringwood 1975, p.268).

According to the model for earlier tholeiitic volcanism (*see* Figures 57 and 76), much of the rock material transformed to eclogite would have been olivine and pyroxene cumulates and gabbros, formed at depth, as well as some thick komatiitic and tholeiitic lavas. The high Cr and Ni content in the calc-alkalic lavas of the Blake River Group (Jolly 1975) and other Archean calc-alkalic basalts and andesites (Hawkesworth and O'Nions 1977; Nesbitt and Sun 1976; Glikson 1976a; Taylor 1979) may be a reflection of these source rocks. A study of the partition coefficients of Ni and Cr between the andesitic magmas and the remaining garnetiferous residual phases is beyond the scope of this report.

Smith (1978) and Ewing (1979) have suggested that the partial melting of eclogite of intermediate

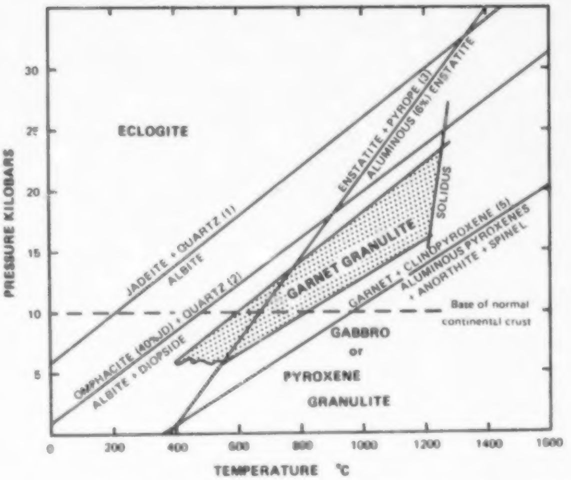


Figure 75. Extrapolated stability fields of eclogite, garnet granulite, and pyroxene granulite-gabbro for the quartz tholeiite (2) composition. The gradient of the eclogite to garnet granulite boundary is 20 bars/°C, and the gradient of the gabbro to garnet granulite boundary is 15 bars/°C (after Green 1975, p.25).

composition was required to produce the REE patterns found in Archean rhyolitic rocks. Because quartz-porphyry rhyolite is found near the base of the Blake River Group, it is unlikely that the succession of calc-alkalic andesitic lavas of the Blake River Group was thick enough so that they could have been transformed into eclogite, and after partial melting, produce rhyolitic magmas. Instead, the rhyolites must have resulted from the fractionation of garnet from andesitic magma at depths of about 100 km (Green and Ringwood 1968). Therefore, the 2 separate populations of points on Figure 71 were probably the result of 2 distinct types of magmas being transported to the surface (Figure 77).

The formation of eclogite would first occur near the centre of the basin of volcanic deposition at its thickest and lowest point. The formation of eclogite would cause an even greater rate of sinking of the basin, particularly in its core. As the eclogite underwent partial melting, andesitic magmas would rise to the surface and replace the tholeiitic magmas as the main contributor to the volcanic deposition in the basin. This would probably result in eclogite forming over an even wider area on the floor of the basin of volcanic deposition, thus allowing calc-alkalic volcanism to become more widespread in the central parts of the basin and also allowing down-buckling to occur over an even wider part of the basin.

Rhyolite and dacite appears to form a high proportion of the younger calc-alkalic rocks toward the center of the basin of volcanism (Figure 65). This would suggest that chambers may have formed in which andesitic magmas were able to fractionate garnet after they became separated from their eclogitic sources. The formation of magma chambers may have been caused by the earlier residual phases of the partially melted eclogite sinking farther into the mantle. As more and more residual material formed, it would sink at an even greater rate than the overlying unmelted eclogite which would have allowed magma chambers to form, and rhyolitic magmas to develop.

THE BLAKE RIVER GROUP IN QUEBEC

The Blake River Group extends from Ontario into Quebec (*see* Figure 64) where it appears to have been deposited in a basin which formed a continuation of the one in Ontario. The Noranda Section of Blake River Group is largely contained between the Destor - Porcupine Fault Zone to the north and the continuation of the Kirkland Lake - Larder Lake Fault Zone (Cadillac, Break, Latulippe 1976) to the south. Thick tholeiitic (Kinojevis Group) and komatiitic (Malartic Group) volcanic successions occur only outside the area contained by the fault zones, so that the amount of deposition of these older ultramafic to mafic volcanic rocks between the fault zones is uncertain. Based on observations that thick successions of komatiitic and tholeiitic lavas were deposited between the 2 fault zones in Ontario, it is likely that this part of the basin also had similar accumulations.

The calc-alkalic volcanism of the Blake River Group in the Noranda area, besides forming most of the rocks between the 2 fault zones, and being downfaulted so that it became juxtaposed with rocks of the Malartic and Kinojevis Groups has a much greater proportion of rhyolite than in Ontario (*see* Figure 64). As in Ontario, the initial area of calc-alkalic volcanism probably occurred in the centre of a basin of volcanic deposition. From here, volcanism gradually expanded to include most of the area between the 2 fault zones. The thickness of the Blake River Group in Quebec is difficult to measure, since many of the older lavas would be expected to have been buried under succeeding volcanic lavas and tuffs and to have been cut off by the major fault zones; but, judging from the conditions observed in Ontario, the accumulated thickness of the Blake River Group in the Noranda area may have been 20 to 30 km.

Rather than forming rhyolite independently of the trondhjemite, Goldie (1978, 1979a, 1979b) proposed that the rhyolites of Noranda were the surface extrusions of the trondhjemitic magmas that caused the Flavrian and Powell Plutons. Assuming these plutons have the HREE depletion observed in Archean trondhjemites elsewhere, their magmas could have been derived either by the partial melting of garnet-bearing amphibolite or eclogite of intermediate composition (Moorbath 1975a), or by the garnet fractionation of an andesitic magma. Fractionation of amphibole, pyroxene, and plagioclase from an andesitic magma at shallow depths should not cause depletion of HREE. Garnet fractionation of an andesitic magma, however, would result in a rhyolitic magma which could produce both trondhjemite intrusions and rhyolite flows. If this was so, the REE patterns found in the trondhjemitic rocks should be similar to those in the rhyolites as observed by Ewing (1979).

To the author's knowledge, the only bodies that cut the Blake River Group are the Flavrian and Powell Plutons (*see* Figure 64). If these intrusions were related to the rhyolitic volcanism, additional

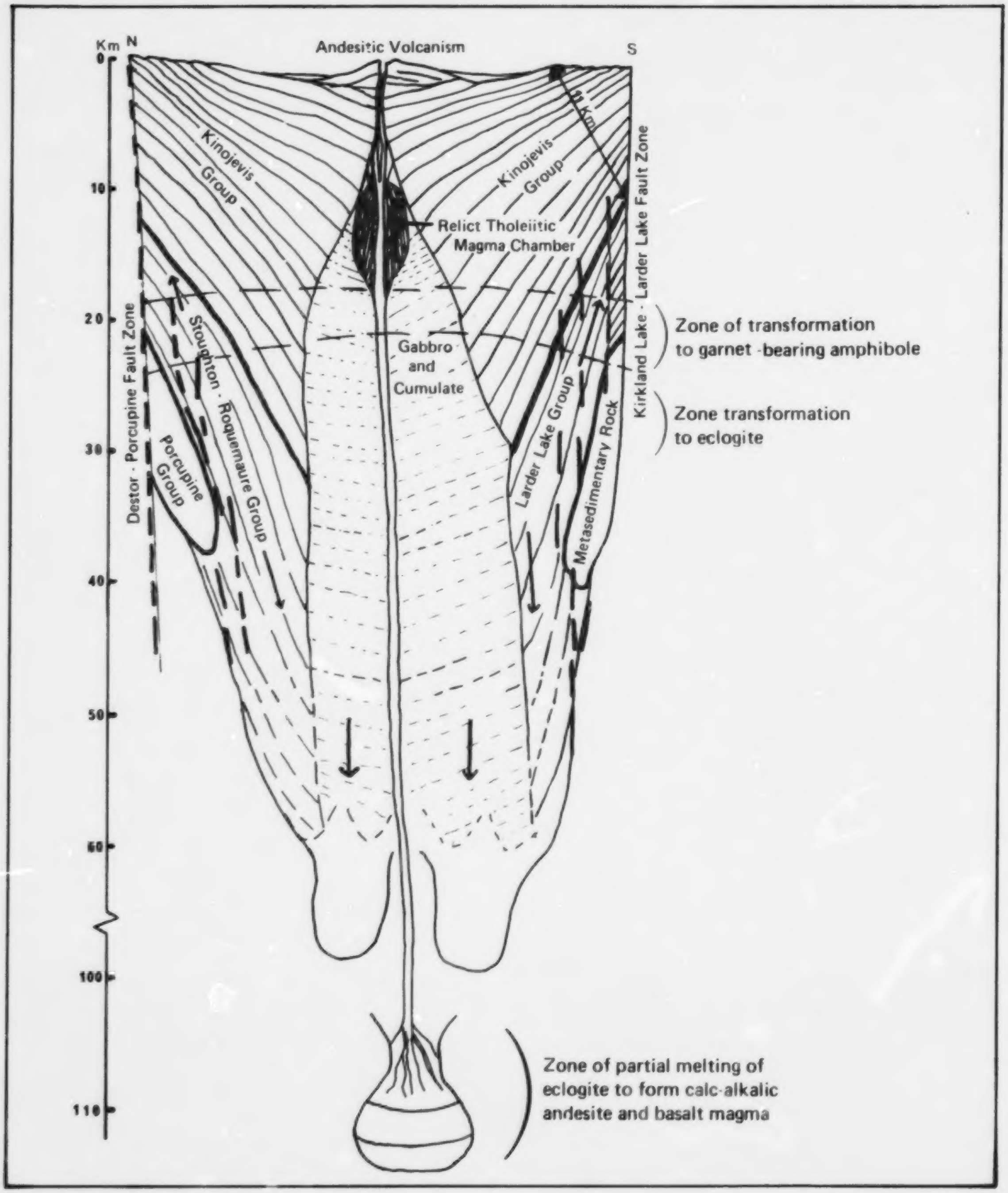

Km
N
Andesitic Volcanism
S
0
10
20
30
40
50
60
100
110
Kinojevis Group
Kinojevis Group
11 Km
Relict Tholeiitic Magma Chamber
Kirkland Lake - Larder Lake Fault Zone
Destor - Porcupine Fault Zone
Stoughton - Roquemaure Group
Porcupine Group
Gabbro and Cumulate
Larder Lake Group
Metasedimentary Rock
Zone of transformation to garnet-bearing amphibole
Zone transformation to eclogite
Zone of partial melting of eclogite to form calc-alkalic andesite and basalt magma

Figure 76. Sketch diagram illustrating the generation of calc-alkalic magmas in the core of the basin of earlier komatiitic and tholeiitic volcanism.

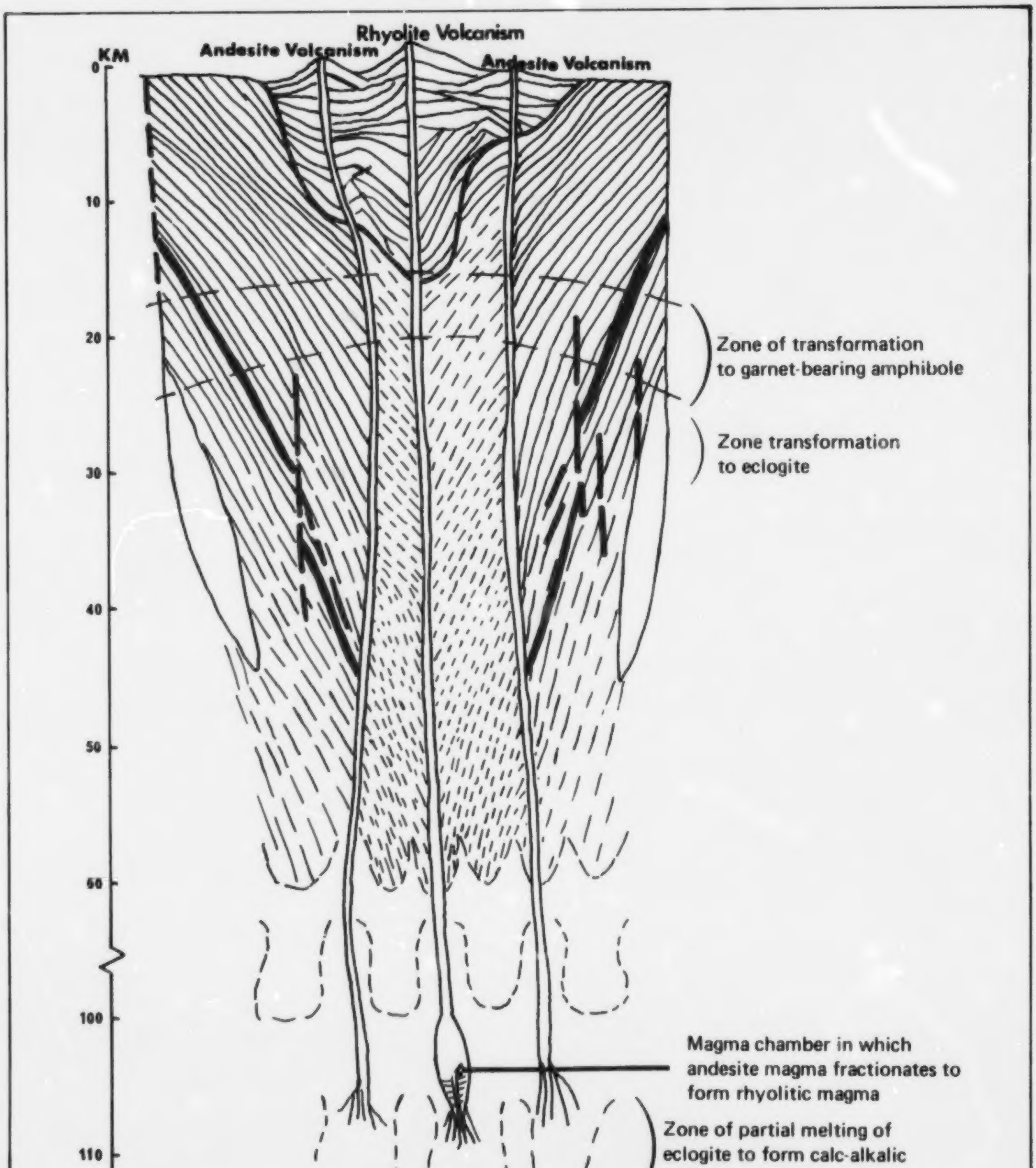

Rhyolite Volcanism
Andesite Volcanism
Andesite Volcanism
KM
0
10
20
30
40
50
60
100
110
Zone of transformation to garnet-bearing amphibole
Zone transformation to eclogite
Magma chamber in which andesite magma fractionates to form rhyolitic magma
Zone of partial melting of eclogite to form calc-alkalic andesite and basalt magma

Figure 77. Sketch diagram illustrating formation of rhyolitic magmas from andesitic magmas of calc-alkalic chemical affinity.

plutons of trondhjemite should be present elsewhere in the Blake River Group to account for the widespread distribution of rhyolite.

The lack of trondhjemite elsewhere in the Blake River Group suggests that calc-alkalic basalt and andesite at its base were not transformed into garnet-bearing granulite and eclogite (N.L. Green 1975) which during further subsidence underwent partial melting, and, by leaving a garnet residue, produced strongly depleted HREE andesitic magma at depths of close to 80 km. Such a magma would fractionate basic plagioclase to form anorthositic rocks in the lower crust (Green and Ringwood 1968) and result in magmas which would be of trondhjemitic composition and able to migrate upward (T.H. Green 1969). In Quebec, the partial melting of calc-alkalic basalt and andesite at depths of 80 km would imply that the earlier formed komatiitic and tholeiitic rocks at the core of the basin in the Noranda area must have been completely displaced by rocks of calc-alkalic chemical affinity. Rhyolitic lavas and trondhjemitic intrusions were also formed independently.

GEOLOGICAL SIGNIFICANCE OF THE BLAKE RIVER GROUP

The Blake River Group represents an arrested stage in the development of sialic crust. The limited generation of trondhjemitic bodies represents a link in the conversion of simatic material to sialic crust that is rarely found or recognized in other greenstone belts. Rhyolite is a subordinate phase of calc-alkalic volcanism.

The key to the preservation of the Blake River and Kinojevis Groups may have been the size of the original basin in which this volcanism took place. As was shown earlier in this report (in the tholeiitic section), the volcanism associated with the deposition of the Kinojevis and Stoughton - Roquemaure Groups north of the Destor - Porcupine Fault Zone was apparently limited by the size of the basin, thus resulting in the formation of the triangular synclinal fold structure. Similarly, the size and shape of the basin south of the Destor - Porcupine Fault Zone in which volcanic rocks of the Upper Supergroup were deposited, may have been limited originally by the location of the earlier formed volcanic piles of the lower supergroups.

Cessation of the tholeiitic volcanism occurred at or shortly after the initiation of calc-alkalic volcanism. This cessation of tholeiitic volcanism was caused possibly by the subsidence of eclogite to mantle depths which closed off the conduits of the tholeiitic magma generated in the mantle, and at the same time, absorbed the heat energy from the mantle so that it became partially melted itself, thus providing a source for the calc-alkalic magmas.

The cessation of calc-alkalic volcanism could have been caused by the exhaustion of parental simatic crustal material available for the generation of calc-alkalic magmas. The amount of partial melting of the eclogite at depth is uncertain. If the partial melting of eclogite was less than 20%, an exhaustion of the parental simatic crustal source would have occurred before the Blake River Group became sufficiently thick to undergo extensive transformation to trondhjemite. A basin of greater lateral extent than that believed by the authors to have existed, would have allowed a larger volume of komatiitic and tholeiitic lava to have accumulated. Such a basin would possibly have provided a greater source of parental material for the calc-alkalic volcanism and ultimately for the formation of trondhjemites and tonalite.

Moorbath (1975a) pointed out that the Sr and Rb isotope data on Archean felsic rocks required that they be generated from simatic crustal material within 50 to 150 Ma after its derivation from the mantle. The zircon U-Pb age dates on the Hunter Mine Group of 2710 ± 2 Ma, and on the Blake River Group in Ontario and Quebec of 2703 ± 2 Ma (Nunes and Jensen 1980) would indicate a large part of the simal to sial transformation occurred within 7 ± 4 Ma in this part of the Abitibi Belt.

The Blake River Group, as a calc-alkalic volcanic pile, does not appear to have reached the same stage in its development as had the calc-alkalic successions of the Lower Supergroups previously discussed. The rocks which remain of the Hunter Mine Group, Upper Formation of the Deloro Group, Skead Group, and Pacaud Tuffs are largely felsic tuffs and tuff-breccias that appear to have been deposited near the edges of volcanic piles. As a consequence, the felsic tuffs are interbedded with iron formation and show a transition from a volcanic to a marine environment (*see* Figure 27). In the Blake River Group, there is a lack of iron formation and well-sorted marine sediments, although Dimroth and Rocheleau (1979), suggest that a succession of sedimentary rocks was starting to be deposited on the edges of the Blake River Group in the Noranda area.

Most of the volcanic rocks in the Blake River Group are lava flows of calc-alkalic basalt and andesite with some flows of dacite and rhyolite, whereas the rocks in the older groups are fragmental units of andesite, dacite, and rhyolite. Some pillowed calc-alkalic andesite lavas are in parts of the Hunter Mine Formation and the Upper Formation, of the Deloro and Skead Groups.

The intrusive rocks in the Blake River Group are mainly stocks and sills of gabbro, quartz gabbro, diorite, and quartz diorite and massive bodies of dacite and rhyolite which can be related to the calc-alkalic igneous activity. In the older calc-alkalic successions, the intrusions are mainly swarms of dacite and rhyolite feldspar and quartz porphyry dikes which intrude each other as well as the fragmental felsic volcanic rocks to form rhyolitic complexes. In addition, sills, and plugs of peridotite, dunite, and iron-rich gabbro intrude the tops of the lower calc-alkalic successions. These intrusions are thought to have marked the initiation of a new volcanic cycle, the upper part of which is considered to be represented by the Blake River Group. Ultramafic intrusive rocks are not observed to intrude the Blake River Group, which suggests that a new volcanic cycle was not initiated in the Kirkland Lake part of the Abitibi Belt following the calc-alkalic volcanism of the Blake River Group.

Large tonalite and trondhjemite batholiths intrude the base of the Hunter Mine Group and the Pacaud Tuffs (*see* Figure 5). The transition from a volcanic to a marine environment suggests that the main centres

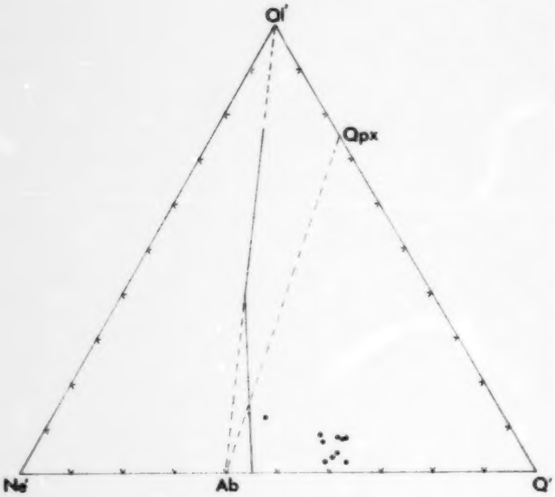


Figure 78. Ol'-Ne'-Q' projections of chemical analyses from the Lake Abitibi Batholith.

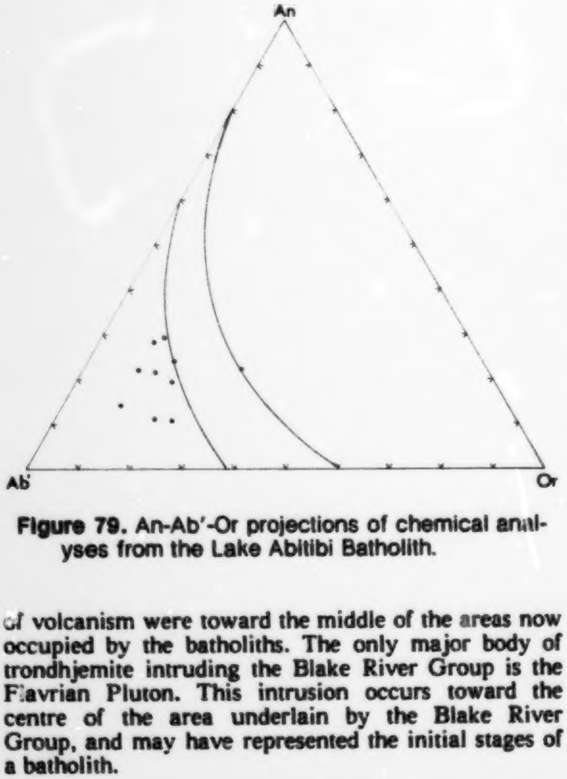


Figure 79. An-Ab'-Or projections of chemical analyses from the Lake Abitibi Batholith.

of volcanism were toward the middle of the areas now occupied by the batholiths. The only major body of trondhjemite intruding the Blake River Group is the Flavrian Pluton. This intrusion occurs toward the centre of the area underlain by the Blake River Group, and may have represented the initial stages of a batholith.

BATHOLITHS OF NA–RICH GRANITIC ROCKS

The Lake Abitibi, Kenogamissi, and Round Lake Batholiths form large areas of Na-rich granitic rock around the periphery of the synclinorium (*see* Figure 5). The Lake Abitibi Batholith, the southern part of which occurs in the Lightning River map area (Map P.2433, back pocket), illustrates the processes by which much, or all of the calc-alkalic rock of a volcanic pile was transferred to Na-rich granitic rocks to form a granitic batholith. If the Round Lake Batholith has features such as REE distribution which are similar to the Lake Abitibi Batholith, it too probably represents a calc-alkalic volcanic pile transformed into a batholith. Until such features as REE distribution are measured, this a remains speculative statement.

LAKE ABITIBI BATHOLITH

Description

The Lake Abitibi Batholith is about 30 by 50 km in area. It has a few elliptical lobes which extend into the surrounding volcanic rocks, and a few smaller satellite bodies occur in the nearby rocks. In many places, the wall rocks were metamorphosed to amphibolite facies, but grade into zones of greenschist and subgreenschist metamorphic facies within 1000 m of the batholith. This indicates that the batholith was emplaced at depths of 2 to 3 km.

The batholith is a composite intrusion with several bodies of massive trondhjemite up to 6 km across, surrounded by gneissic trondhjemite and tonalite as well as minor quartz-diorite (*see* Figures 23 and 81). The bodies of massive trondhjemite grade into the gneissic rocks such that the gneissic rocks appear to form halos around the massive bodies (Map 2433, back pocket). In each halo, the gneissosity dips away from the central massive body, suggesting vertical intrusion as a mechanism of emplacement.

The massive bodies are homogeneous, whereas the gneissic halos tend to contain several phases of gneissic trondhjemite and tonalite having different colour, grain size, and proportions of mafic and felsic minerals. The gneissic phases cut and grade into one another, but all have gneissosities indicating similar strikes and dips. The gneissic textures consist of altered plagioclase augen altered and weakly deformed, surrounded by recrystallized, finer grained, interlocking quartz and albite that form oval volumes enclosed by biotite. This would suggest that the gneissic textures were developed largely through recrystallization during deformation as the late phases of trondhjemitic magmas domed the surrounding rocks.

Trondhjemitic composition pegmatites with both gneissic and massive textures, cut the gneissic rocks. These dikes represent fracture fillings since the elongate plagioclase and quartz crystals extend perpendicularly to the walls of the dikes (*see* Figure 18). Pegmatite dikes are rare in the massive trondhjemite bodies.

Chemical analyses of 10 specimens collected from widely separated locations showed that although the samples varied from strongly gneissic to massive, from very coarse grained to medium-grained rocks, the chemical compositions were almost the same. From Ol'-Ne'-Q'and An-Ab'-Or diagrams (Figures 78 and 79) the rocks are quartz-normative and orthoclase-poor and on the Jensen Cation Diagram (Figure 80), the rocks are equivalent to calc-alkalic dacite and rhyolite. The limited chemical variation and composition of the rocks suggest that near-surface assimilation of rocks, more mafic than calc-alkalic andesite, probably was not significant.

The Lake Abitibi Batholith (Figure 81) intrudes komatiitic and tholeiitic lavas belonging to the Stoughton - Roquemaure Group as well as the calc-alkalic rocks of the Hunter Mine Group, but exposures of the contact of these rocks with the batholith are rarely observed, except where they form inclusions in granitic rocks. Within 100 m of the batholith, the komatiitic and tholeiitic lavas are amphibolitized and schistose, and the pillows have an elongation several times their width. The original selvages and tails of the pillows are well preserved, suggesting that they mainly underwent plastic deformation.

At the granite contacts, the komatiitic and tholeiitic lavas have quartz and feldspar prophyroblasts, 2 to 4 mm in size and take on a gneissic texture. The contact is gradational within the width of 1 or 2 m, implying that the assimilation of the mafic volcanic rocks was limited.

Inclusions of amphibolitized mafic volcanic rocks, 0.5 to 5 m in size, within marginal granitic rocks show varying degrees of assimilation. Most xenoliths are within 1 km of the contact, but some, up to 5 km inside the batholith, may represent pieces of roof rocks. If large amounts of mafic and ultramafic rock had become assimilated, the rocks of the batholith should show a much greater variation in chemical composition. If large amounts of calc-alkalic dacite and rhyolite were incorporated, little reaction with the magma would have been expected, but some of the 0.5 to 1 mm grained, grey gneissic tonalite may represent assimilated blocks of felsic calc-alkalic rock incorporated into the batholith.

Petrogenesis

If the trondhjemites and tonalites of the Lake Abitibi Batholith are strongly depleted in HREE, as they appear to be in other Archean terrains, then such rocks were formed from magmas derived by partial melting of calc-alkalic rocks during which the garnet remained in the residue. The trondhjemites and tonalites are younger than the komatiitic and tholeiitic lavas of the Stoughton - Roquemaure Group, and thus postdate the calc-alkalic Hunter Mine Group. In addition, there are few calc-alkalic rocks above the Hunter Mine Group that can be related to development of the batholith. This suggests that the magmas were derived from the partial melting of Hunter Mine calc-alkalic basalts and andesites, which were partially melted to eclogite.

The Hunter Mine Group probably once represented a thick calc-alkalic volcanic pile that may have had a trondhjemitic core (Figure 82a) similar to the Flavarin Pluton in the younger Blake River Group. Conversion of the lower calc-alkalic basalts and andesites to eclogite would likely have continued, and additional trondhjemitic magmas would have migrated upward (Figure 82b). Once emplaced near surface, trondhjemitic intrusions, because of their low density, would not likely subside to compensate the sinking eclogite at depth. Instead, additional calc-alkalic basalts and andesite of the calc-alkalic pile would be more likely to subside downward and inward to follow the sinking eclogite. This process would continue if denser komatiitic and tholeiitic lavas were deposited on top of the calc-alkalic lavas marginal to the main calc-alkalic volcanic pile, as shown in Figure 57. As the mafic lavas accumulated and tilted, the underlying calc-alkalic volcanic rocks would continue to gravitate downward and inward (Figure 82b). This would alleviate the problem of displacing the calc-alkalic rocks below the mafic lavas to provide additional room for succeeding mafic lava and, at the same time, would remove most of the evidence of a pre-existing calc-alkalic volcanic pile. In addition, xenoliths of komatiitic and tholeiitic lavas would be expected to be found in the batholith as the trondhjemitic magmas migrated upward and intruded the komatiitic and tholeiitic lavas near surface. The lack of calc-alkalic volcanic rocks among the komatiitic and tholeiitic lavas also suggest that the batholith was emplaced without associated calc-alkalic volcanism.

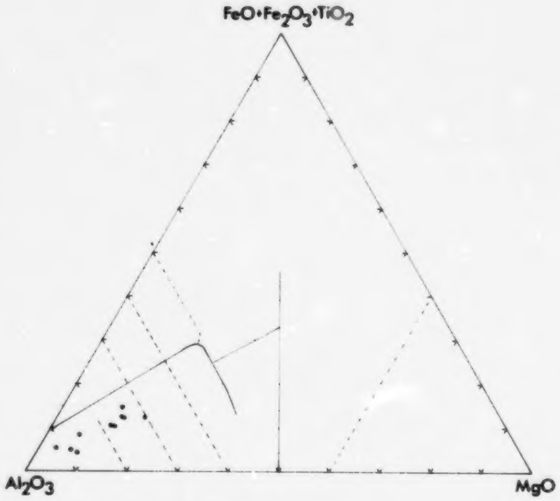


Figure 80. Chemical analyses of the Lake Abitibi Batholith projected on the Jensen Cation Diagram.

ROUND LAKE BATHOLITH

The Round Lake Batholith south of the Kirkland Lake - Larder Lake Fault Zone forms an elliptical area 30 km by 80 km and, except for its eastern side, contacts with the volcanic rocks are obscured by younger Proterozoic sedimentary and intrusive rocks (*see* Figure 5). This batholith is a composite intrusion composed mainly of massive and gneissic tonalite and trondhjemite. In addition, some potassium-rich granitic intrusive bodies are present in its northern edge. Like the Lake Abitibi Batholith, the Round Lake Batholith is surrounded by subgreenschist facies metamorphic volcanic rocks (Jolly 1978). This suggests that the batholith was emplaced at shallow depths.

Except for a narrow zone of calc-alkalic as Pacaud Tuffs, no evidence of major calc-alkalic volcanism remains. However, the Pacaud Tuffs appear to represent the transition zone between a volcanic and a marine environment, with major calc-alkalic volcanism probably in the area now occupied by the Round Lake Batholith.

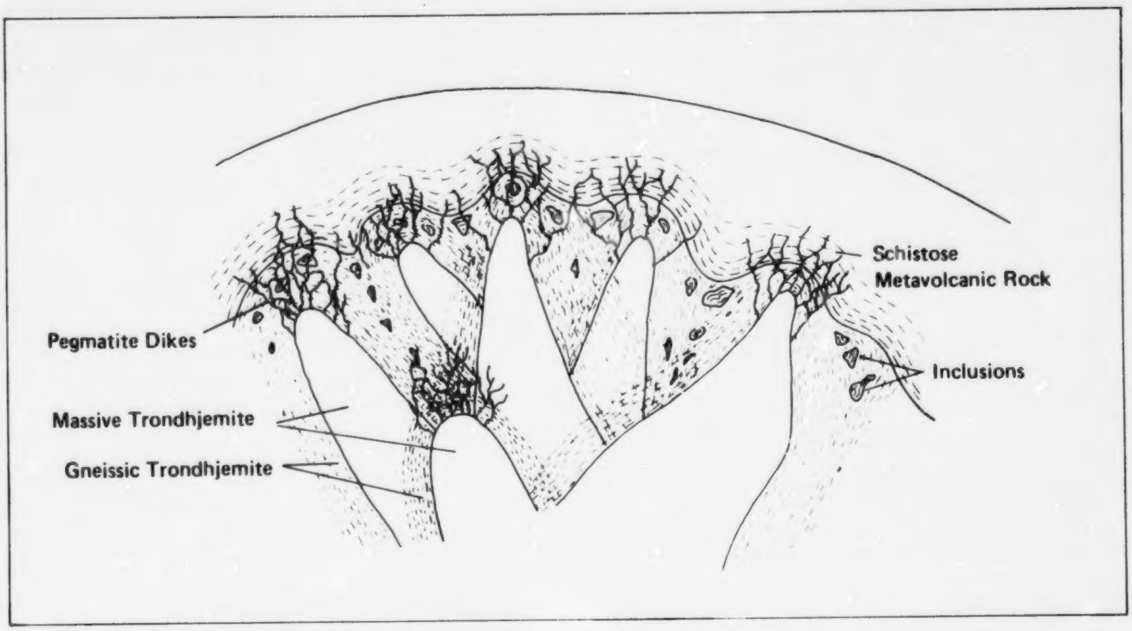


Figure 81. Cross-sectional sketch diagram of the rocks in the Lake Abitibi Batholith.

The apron of calc-alkalic volcanic rocks and the sedimentary rocks farther out, were likely displaced down toward the root of the Round Lake Batholith by the rocks belonging to the Wabewawa, Catherine, and Skead Groups which now form monoclinal volcanic sequence north, east, and south of it (*see* Figures 5 and 63).

ALKALIC VOLCANIC SUCCESSIONS AND K–RICH GRANITIC INTRUSIONS

Except for late diabase dikes, the last igneous events in the Kirkland Lake part of the Abitibi Belt were the deposition of the alkalic volcanic rocks and the intrusion of K-rich granitic rocks. In the Lightning River - Magusi River map area, the alkalic volcanic rocks with their associated sedimentary and intrusive rocks around Garrison Township (Figure 83) are referred to as the Destor - Porcupine Complex. These rocks are poorly exposed, and most of the information comes from drill core data, so that only a limited discussion of these rocks is possible in this report. Better exposed alkalic volcanic rocks occur in the Timiskaming Group near Kirkland Lake (Cooke 1966). Here, the volcanic rocks are interlayered with fluvial sedimentary rocks (Hyde 1978, 1980), and are cut by alkali-rich dikes and stocks. The remaining rocks include the Watabeag Batholith and numerous other K-rich granitic stocks shown on Figure 83.

Because these rocks unconformably overlie and cut all other rocks, they represent the final stages of the volcanic cycle which formed the Upper Supergroup (Cooke 1966; Goodwin and Ridler 1970). However, unless the magma sources for these rocks were associated with the development of the Upper Supergroup, the alkalic rocks should not be accepted as being a genetic part of this supergroup.

DESTOR - PORCUPINE COMPLEX

Volcanic Rocks

The volcanic rocks of the Destor - Porcupine Complex are restricted to the area between the middle and south branches of the Destor - Porcupine Fault Zone (*see* Figure 19). In this area, these rocks are contained in fault-bounded wedges juxtaposed against similar wedges of komatiitic and Mg-rich tholeiitic basalt lavas belonging to the Stoughton - Roquemaure Group. Within the wedges, the alkalic volcanic rocks occur interlayered with sedimentary rocks (Figure 84). The wedges of rock belonging to the Destor - Porcupine Complex and to the Stoughton - Roquemaure Group together, occupy an area 1 to 2 km wide and more than 30 km long between the 2 branches of the Destor - Porcupine Fault Zone. This area extends west through Holloway, Harker, and Garrison Townships and continues westward (*see* Figure 2). Farther west, in Hislop Township, Prest (1957) observed that the alkalic volcanic rocks and their associated sedimentary rocks were juxtaposed, by downfaulting along a northeast-trending fault, against mafic lavas referred to in this study as the Kinojevis Group. This would suggest that the wedges of rocks in the Lightning River - Magusi River area, were also preserved by downfaulting along the Destor - Porcupine Fault Zone.

Major element analyses were made on 37 specimens of rocks that were interlayered with sediments, had textures preserved that are indicative of lava flows or tuffs, and were as far as possible from known felsic intrusions.

On the Ol'-Ne'-Q'diagram (Figure 85), unlike previously described groups of rocks, the analyses of the alkalic volcanic rocks fall mainly in the lower half of the alkalic field near the Ol'-Ab join. Some are

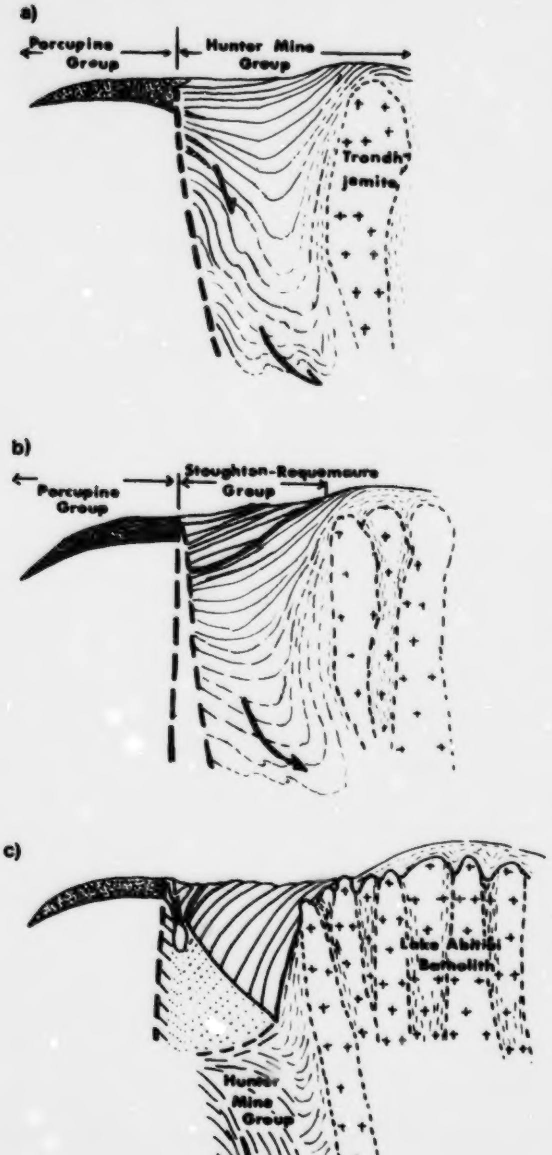


Figure 82. Diagram of the deposition of the Stoughton - Roquemaure Group adjacent to the developing Lake Abitibi Batholith: a. The calc-alkalic pile represented by the Porcupine and Hunter Mine Groups prior to the deposition of the Stoughton - Roquemaure Group (*see* Figures 23, 28, and 57); b. Deposition of the Stoughton - Roquemaure Group on the edge of the calc-alkalic pile as the Lake Abitibi Batholith developed in the central parts of the calc-alkalic pile; c. Downward displacement of the Hunter Mine Group and the growth of the Lake Abitibi Batholith to intrude the Stoughton - Roquemaure Group.

toward the Ne'apex and classify as nephelinites (Irvine and Baragar 1971; Figure 86). Five of the rocks have normative quartz and have the chemical composition of calc-alkalic dacite.

The An-Ab'-Or diagram (Figure 87) indicates that the alkalic volcanic rocks are part of the sodic series and range in composition from alkalic basalt to trachyte (Figure 86).

The analyses of the alkalic volcanic rocks plotted on the Jensen Cation Diagram (Figure 88) form 2 distinct clusters; one in the iron-rich tholeiitic basalt field and the other in the calc-alkalic andesite dacite and rhyolite fields. Some of the analyses which project in the iron-rich tholeiitic field are titanium-rich basalts from drill holes of Anatole Resources Incorporated in Harker Township (*see* Table 4) which are interlayered with rocks whose chemical compositions project onto the calc-alkalic field and have less than 0.5% TiO_2. The alkalic volcanic rocks may have been formed by 2 distinct magmas, 1 generated by a low degree of partial melting of the mantle and the other by partial melting of parental crustal rocks. Trace element studies should be measured to investigate these possibilities.

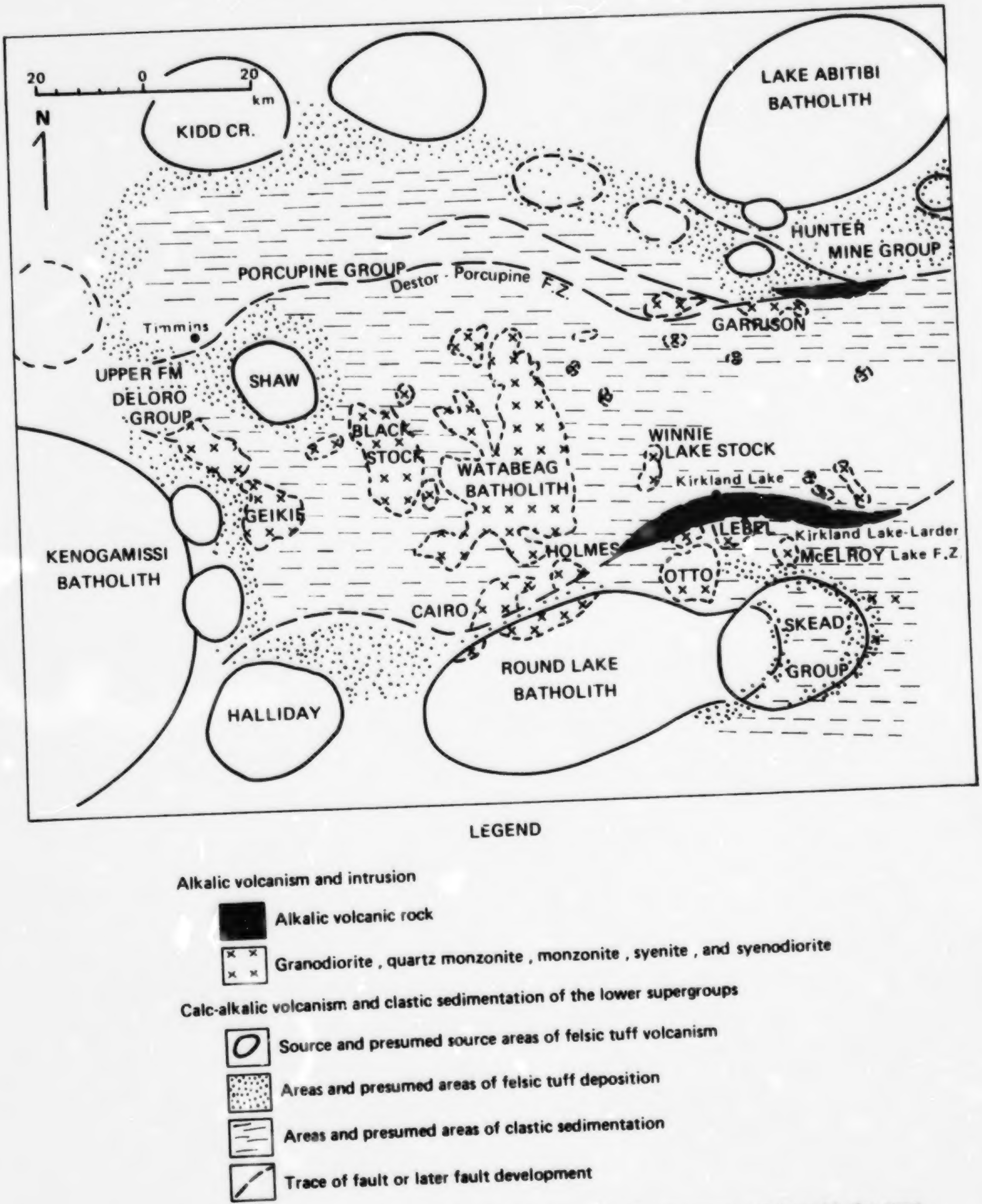
Sedimentary Rocks

The sedimentary rocks of the Destor - Porcupine Complex differ from those of the Porcupine Group by the presence of conglomerates which contain rounded clasts of alkalic volcanic rock, mostly trachyte, and rounded clasts of syenodiorite, granodiorite, and monzonite that range from porphyritic to equigranular in texture. A few clasts of chert, jasper, iron formation, and mafic to ultramafic rock derived from komatiitic and tholeiitic lavas are also present.

Satterly (1954) suggested that the conglomerates in Holloway Township might be Timiskaming-type rocks. Farther west, zones of similar conglomerate were identified in Guibord Township (Prest 1953, p.10-23) and Hislop Township (Prest 1957). Diamond-drill holes have intersected these conglomerates, with intercalated alkalic volcanic rocks, in Harker and Garrison Townships along the southern branch of the Destor - Porcupine Fault Zone.

The clasts are supported in a matrix composed of rounded feldspar and quartz grains, and rock fragments cemented by quartz, albite, chlorite, calcite, and sericite. Walker (1976) suggested that the conglomerates were deposited by turbidity currents. Because of the limited exposure of these rocks, little else is known about their environment of deposition. On the basis of the clasts, most of the rocks were derived from the flows and tuffs interlayered with them. The clasts of syenodioritic and monzonite were derived from dikes which cut these rocks. The clasts of komatiite and tholeiite were probably derived from the Stoughton - Roquemaure Group and the chert and iron formation from the Hunter Mine Group. The presence of many clasts of local sources, suggests that the rocks were deposited in a graben or narrow trough formed along the Destor - Porcupine Fault Zone.

Thin beds of greywacke and shale occur interlayered with the conglomerate and alkalic volcanic rocks. These beds are composed of numerous graded beds, 1 to 30 cm thick, and tops mainly face south.


20
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km
N
KIDD CR.
LAKE ABITIBI BATHOLITH
HUNTER MINE GROUP
PORCUPINE GROUP
Destor - Porcupine F.Z.
Timmins
GARRISON
UPPER FM
DELORO GROUP
SHAW
BLACK STOCK
WATABEAG BATHOLITH
WINNIE LAKE STOCK
Kirkland Lake
GEIKIE
KENOGAMISSI BATHOLITH
HOLMES
LEBEL
Kirkland Lake-Larder
McELROY Lake F.Z
OTTO
CAIRO
SKEAD GROUP
ROUND LAKE BATHOLITH
HALLIDAY
LEGEND
Alkalic volcanism and intrusion
Alkalic volcanic rock
Granodiorite, quartz monzonite, monzonite, syenite, and syenodiorite
Calc-alkalic volcanism and clastic sedimentation of the lower supergroups
Source and presumed source areas of felsic tuff volcanism
Areas and presumed areas of felsic tuff deposition
Areas and presumed areas of clastic sedimentation
Trace of fault or later fault development

Figure 83. Map of the alkalic volcanism and K-rich intrusions in the Timmins - Kirkland Lake area.

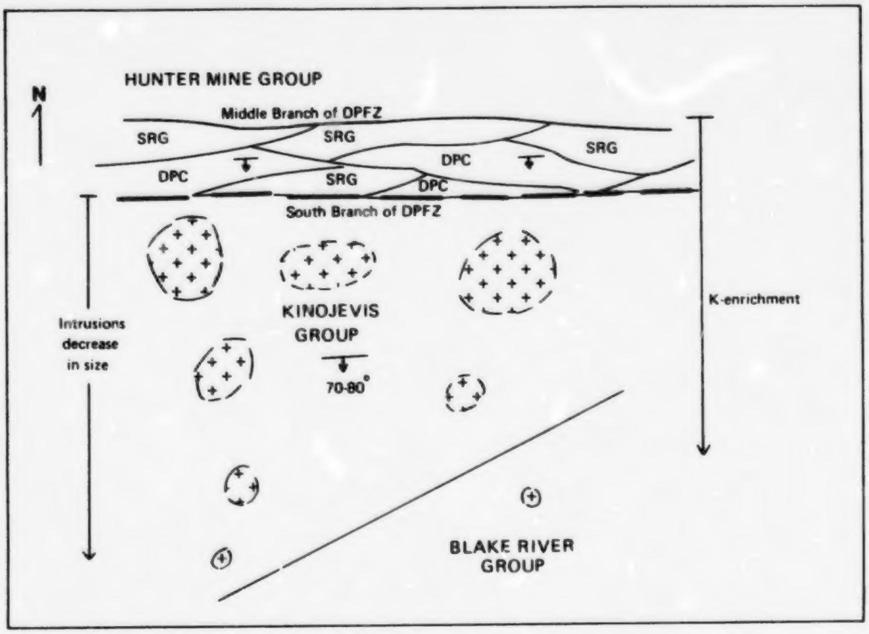


Figure 84. Diagram of the Destor - Porcupine Complex. Fault-bounded wedges of rocks of the Stoughton - Roquemaure Group (SRG) and of the Destor - Porcupine Complex (DPC) occur between the Middle and South branches of the Destor - Porcupine Fault Zone (DPGZ).

Intrusive Rocks

Syenodiorite, monzonite, quartz monzonite granodiorite, and minor syenite form the intrusive rocks of the Destor - Porcupine Complex. The syenodiorites, which are mainly dikes, cut the stratiform rocks of the complex as well as those of the Stoughton - Roquemaure Group between the 2 branches of the Destor - Porcupine Fault Zone. These syenodiorite dikes may have been feeders for the trachytic volcanic rocks. South of the fault zone, the intrusive rocks are mainly stocks composed of granodiorite and quartz monzonite. Farther to the south, the stocks which decrease in size and number are mainly monzonite and syenite. This indicates that there is a southward potassium-enrichment in the intrusive rocks. North of the middle branch of the Destor - Porcupine Fault Zone, the absence of dikes and stocks of the Destor - Porcupine Complex suggests that the distribution of the intrusive rocks, like that of the alkalic volcanic and sedimentary rocks, was partly fault controlled.

Unlike the trondhjemitic batholiths, the intrusions associated with the Destor - Porcupine Complex have aureoles of K, Na, CO_2 and H_2O metasomatism several times their size. These aureoles occur particularly in the rocks of the Stoughton - Roquemaure Group between the middle and south branches of the Destor - Porcupine Fault Zone (*see* Figures 46 and 47). This strong metasomatism may indicate that the magmas either came from sources rich in volatiles and water, or became enriched *en route* to the surface.

Many of the smaller intrusions of the Destor - Porcupine Complex appear to have reacted with the mafic volcanic rocks resulting in hybrid rock which can form up to 50% of the volume of each intrusion and leaving only the core having a felsic composition. In places, the hybrid rocks grade into lamprophyres. Pyroxene and hornblende hornfels facies metamorphism occurs near the contact of some of the larger stocks, and suggest a temperature of 700 to 800°C occurred during metamorphism (Turner 1968).

Of the plots on the Ol'-Ne'-Q'diagram (Figure 89), those which lie close to the Ab point, are mainly of syenodiorite and syenite dikes, and those with normative quartz are from granodiorite and quartz monzonite stocks found in Garrison and Harker Townships.

On the An-Ab'-Or diagram (Figure 90), the orthoclase-poor analyses are mainly from dikes of syenodiorite and the remainder of the rocks are granodiorites, monzonites, quartz-monzonites, and syenites except for 2 analyses which are of hybrid rocks and have a high An normative content. The Jensen Cation Diagram (Figure 91), shows the intrusive rocks are calc-alkalic dacite and rhyolite, except for the hybrid rocks which are equivalent calc-alkalic andesite and basalt.

Petrogenesis

South of the middle branch, the sedimentary rocks and calc-alkalic tuffs marginal to the calc-alkalic volcanic pile (*see* Figure 27) represented by the Hunter Mine and Porcupine Groups must have been displaced downward 20 to 40 km in order to accommodate the overlying rocks of the Stoughton - Roquemaure and Kinojevis Groups (*see* Figure 76). The intrusions of the Destor - Porcupine Complex occur up to 15 km south of the fault zone. This suggests that the parental material for their magmas in this area may have been downfaulted marginal water-rich calc-alkalic tuffs and sedimentary rocks of the Hunter Mine and Porcupine Groups and the ultramafic lavas of the Stoughton - Roquemaure Group.

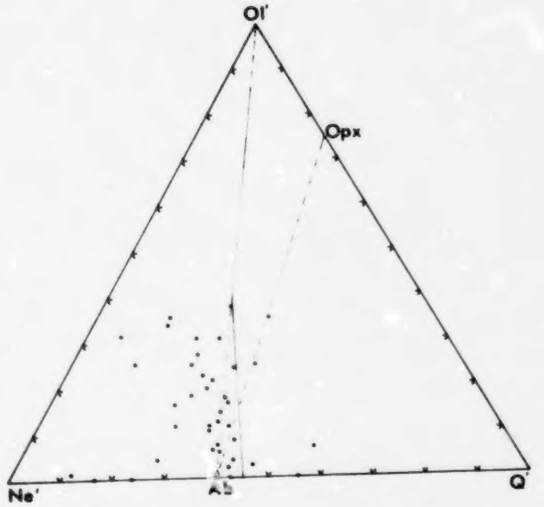


Figure 85. Ol'-Ne'-Q' projections of chemical analyses of alkalic volcanic rock of the Destor - Porcupine Complex.

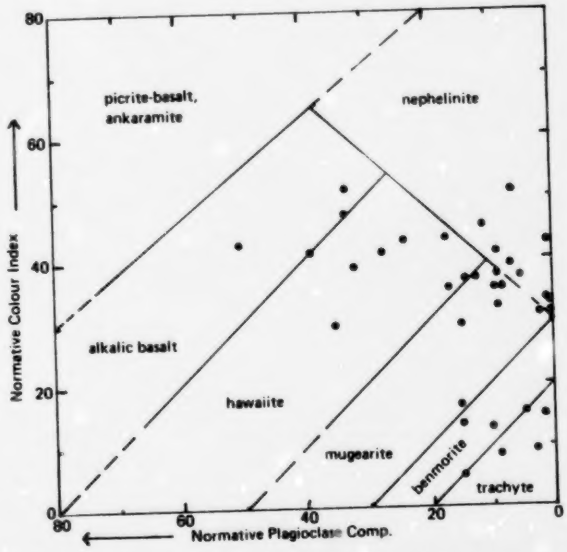


Figure 86. Normative Colour Index versus Normative Plagioclase Composition (molecular) of chemical analyses of the alkalic volcanic rocks from the Destor - Porcupine Complex (rock type-field boundaries for Na-series alkalic volcanic rocks after Irvine and Baragar 1971).

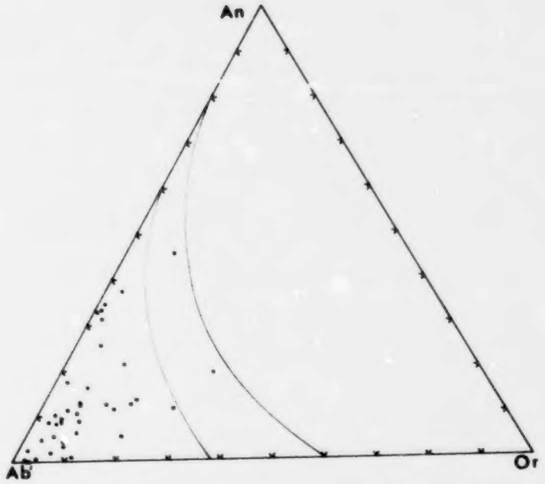


Figure 87. An-Ab'-Or projections of chemical analyses of alkalic volcanic rocks of the Destor - Porcupine Complex.

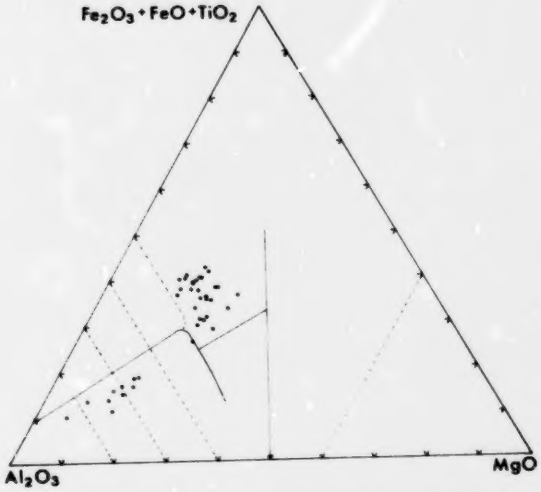


Figure 88. Chemical analyses of the alkalic volcanic rocks of the Destor - Porcupine Complex projected on the Jensen Cation Diagram.

At depths of 30 to 40 km, these rocks would have been transformed into garnet-bearing amphibolite or eclogite and could be similar sources to those for the quartz monzonites and syenites of northeastern Minnesota as proposed by Arth and Hansen (1975). The thickness of these rocks would be expected to decrease southward, concurrent with the observation that the stocks decrease in size southward. The increase in K_2O in these rocks may have been a function either of the depth, or of the composition of the rocks from which the magmas were generated. The sedimentary rocks to the south would be expected to be buried more deeply and to contain a higher proportion of argillite in which K_2O would be more abundant than in the calc-alkalic tuffs to the north (Ronov and Yaroshevskiy 1976).

The wedges of alkalic volcanic rock along the southern part of the Destor - Porcupine Fault Zone suggest that the fault zone became a graben along which these wedges could be downthrown and preserved. Dilation of the fault zone may have occurred as the volcanic rocks to the south subsided and collapsed southward because of the sinking of eclogite,

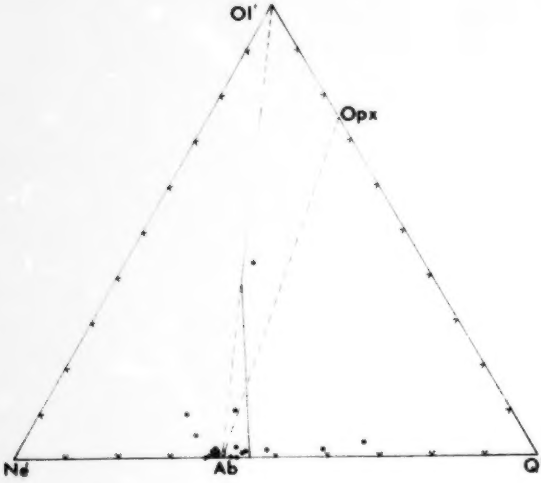


Figure 89. Ol'-Ne'-Q' projections of the chemical analyses of intrusive rocks of the Destor - Porcupine Complex.

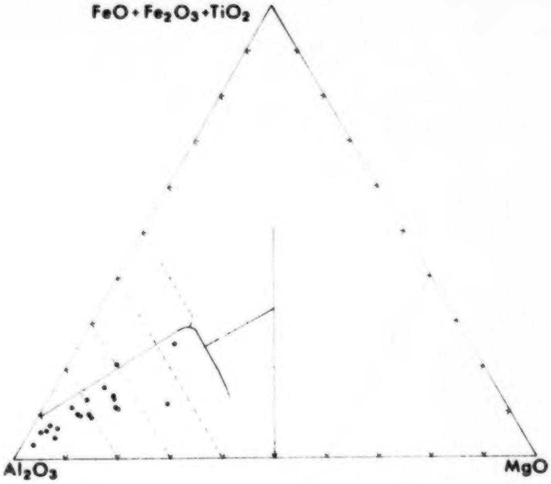


Figure 91. Projections of the chemical analyses of intrusive rocks of the Destor - Porcupine Complex on the Jensen Cation Diagram.

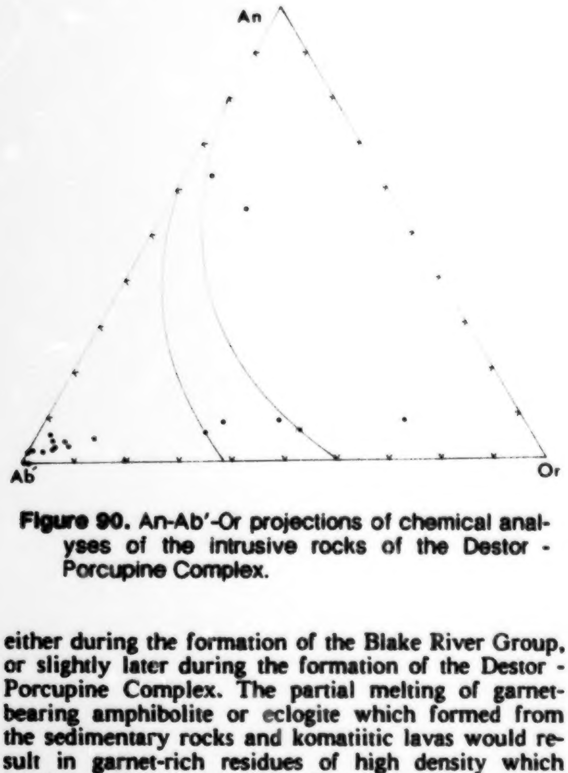


Figure 90. An-Ab'-Or projections of chemical analyses of the intrusive rocks of the Destor - Porcupine Complex.

either during the formation of the Blake River Group, or slightly later during the formation of the Destor - Porcupine Complex. The partial melting of garnet-bearing amphibolite or eclogite which formed from the sedimentary rocks and komatiitic lavas would result in garnet-rich residues of high density which would continue to subside. The release in pressure along the Destor - Porcupine Fault Zone may have allowed magmas to penetrate upward so that alkalic volcanism occurred.

The syenodiorite intrusions and the felsic sodic-series alkalic to calc-alkalic volcanic rocks were probably generated by the partial melting of calc-alkalic tuffs and komatiitic lavas transformed at depth into garnet-bearing amphibolites. The mafic alkalic volcanic rocks, because of their high-TiO_2 contents, may have been generated by small degrees of partial melting of the mantle rocks. Dialation of the fault zone, which extended to depths of 30 to 40 km, would be expected to cause a release in pressure and allow adiabatic melting of the mantle rocks (McKenzie 1969).

The lack of alkalic volcanic rocks and intrusive rocks belonging to the Destor - Porcupine Complex north of the middle branch of the Destor - Porcupine Fault Zone would suggest that the sources for these magmas were not available below the rocks of the calc-alkalic Hunter Mine Group in this area.

WATABEAG BATHOLITH AND NEARBY K-RICH GRANITIC INTRUSIONS

The Watabeag Batholith is a composite intrusion with phases of granodiorite, quartz monzonite, and syenite. The batholith as well as with several other smaller plutons of similar composition, occupies a broad area southeast of Timmins (*see* Figures 5 and 83; Pyke 1970, 1975, 1978a, 1978b, 1980). A thick wedge of sedimentary rock equivalent to the Porcupine Group was probably deposited in this area on the flanks of calc-alkalic volcanic piles to the north, west, and south (*see* Figure 29). These sedimentary rocks may have been buried deeply enough by komatiitic lavas and tholeiitic lavas to have been transformed into garnet-bearing amphibolites and eclogites which underwent partial melting.

THE ROUND LAKE BATHOLITH

Bodies of K-rich granitic rock are observed to cut trondhjemitic batholiths in most Archean terrains (Hunter 1973; Glikson 1979b). An example of this is the northwestern part of the Round Lake Batholith where it is cut by granodiorite, quartz monzonite, and syenite (*see* Figure 83). Here, the batholith intrudes tholeiitic lavas belonging to the Kinojevis Group near the proposed westward extension of the Kirkland Lake - Larder Lake Fault Zone (*see* Figure 5).

In this report, in the section on the trondhjemitic batholiths, the suggestion was made that the outer parts of the pre-existing calc-alkalic piles and associated sedimentary rocks would be displaced downward and inward, toward the core of a batholith being formed (Figure 82) because these rocks were covered and were displaced downward by thick and extensive successions of younger komatiitic and tholeiitic lavas (*see* Figure 76). The sedimentary rocks, because of their greater distance from the developing batholith, would be expected to undergo partial melting later than most of the calc-alkalic rocks. In addition, some ultramafic lavas could also be transformed into garnet-bearing amphibolite and eclogite and undergo partial melting at depth with the sedimentary rocks already converted to eclogite. This would produce the mixed parental ultramafic and sedimentary rock for the syenites, as proposed by Arth and Hanson (1975).

K–RICH IGNEOUS ROCKS NEAR KIRKLAND LAKE

The K-rich igneous rocks near Kirkland Lake occur in a broad area, 25 km wide and 60 km long, north of the Round Lake Batholith (Figure 83). These rocks include the Timiskaming Group volcanic rocks (Cooke 1966), and possibly some trachytic volcanic rocks in the Skead Group (Hewitt 1951), as well as many of the dikes and stocks which cut all the groups of rocks. These include rocks of the Blake River Group, a short distance north of the Kirkland Lake - Larder Lake Fault Zone.

A pebble conglomerate containing syenodiorite and trachyte clasts occurs at the contact between the Skead Group of the Lower Supergroup and the Larder Lake Group of the Upper Supergroup, thus supporting Hewitt's (1951) observation of trachytic volcanic rocks at the top of the Skead Group. This, combined with the Timiskaming Group unconformably overlying the volcanic rocks (Kinojevis and Blake River Groups) north of the Kirkland Lake - Larder Lake Fault Zone (Thomson 1949) indicates 2 periods of alkalic volcanism and intrusion occurred.

Some sedimentary rock postulated to have been deposited earlier on the northern flanks of the calc-alkalic pile represented by the Pacaud Tuffs (Figure 83), could have been partly melted during the subsidence of the Skead Group volcanic rocks. However, the larger, more northerly, remaining part of these rocks probably underwent later subsidence and partial melting which resulted in the formation of the Timiskaming Group and large K-rich granitic stocks.

TIMISKAMING GROUP

The Timiskaming Group occupies an area about 4 km wide and 60 km long that straddles the Kirkland Lake - Larder Lake Fault Zone (Figure 83). On its northern side, the Timiskaming Group unconformably overlies the Kinojevis Group and in places, the Blake River Group (Thomson 1949). Where the Timiskaming Group is nearly horizontal, as at its base, the lava flows of the underlying Kinojevis Group are steeply dipping, face north, and indicate that the attitudes in the rocks to the north were largely developed prior to the Timiskaming Group.

On its southern side, the Timiskaming Group is faulted against the Larder Lake Group (Jensen 1978c). In places, the Timiskaming Group forms wedges of rock faulted in with the Larder Lake Group. The rocks of these wedges as well as the remaining rocks of the Timiskaming Group dip 30 to 70° south.

The Timiskaming Group consists of K-rich alkalic and calc-alkalic volcanic rocks (Cooke 1966) interlayered with fluvial sedimentary rocks (Hyde 1978) and is cut by dikes and stocks of syenite, monzonite, and quartz monzonite (Thomson 1950). The conglomerates of the sedimentary rocks contain alkalic volcanic rock, syenite, and monzonite clasts of local derivation as well as clasts of komatiite, tholeiite, iron formation, and chert which appear to be largely derived from the south. Rock clasts derived from the Kinojevis and Blake River Groups to the north appear to be absent. Because of the low grade of metamorphism, Jolly (1978) concluded that the Timiskaming Group had little erosion, suggesting that it may have been formed, and preserved, in a narrow graben with a scarp on its southern side. The southward dip of the Timiskaming Group would indicate that the scarp formed as the graben subsided along its southern side.

K-RICH GRANITIC INTRUSIONS

K-rich granitic intrusions and syenodiorite cut all the rocks north of the Round Lake Batholith including those in and north of the Kirkland Lake - Larder Lake Fault Zone (*see* Figure 83). The intrusions are very similar to those of the Destor - Porcupine Complex (Thomson 1950; Lawton 1959).

The major intrusions are the Otto, Lebel, and McElroy Stocks which are composed mainly of quartz syenite and quartz monzonite south of the Kirkland Lake - Larder Lake Fault Zone. The intrusions are surrounded by many small dikes and stocks of syenite and monzonite. North of the Kirkland Lake - Larder Lake Fault Zone, the K-rich granitic intrusions become smaller and less numerous. However, there have been no studies to determine if there are any progressive changes in their chemical composition.

PETROGENESIS

A thick wedge of sedimentary rock deposited north of the calc-alkalic pile represented by the Pacaud Tuffs (*see* Figure 23) is suggested as the likely source of the K-rich alkalic volcanic and granitic intrusive rocks near Kirkland Lake. These rocks would have been buried by the Larder Lake Group, and by some rocks associated with the Wabewawa, Catherine, and Skead Groups. Since the Larder Lake Group appears to have been deposited on a shelf south of the Kirkland Lake - Larder Lake Fault Zone, it is unlikely that the earlier sedimentary rocks subsided to any great extent during this period of volcanic deposition. Rather, subsidence at this time probably occurred mainly on the northern side of the developing Kirkland Lake - Larder Lake Fault Zone.

The Timiskaming Group contains K-rich alkalic volcanic rocks, whereas the alkalic volcanic rocks of the Destor - Porcupine Complex are of the Na-series. The K-rich parental sources for the Timiskaming alkalic volcanic rocks may have been largely argillaceous rocks deposited at the edge of a sedimentary shelf, whereas the Na-rich rocks of the Destor - Porcupine Complex may have had parental calc-alkalic tuffs. This would concur with the observation that the greatest volumes of K-rich granitic intrusive rock occur south of the alkalic volcanic rocks in both areas, where the greatest accumulations of clastic sediments were expected to have formed (*see* Figure 23).

The alkalic volcanic rocks and K-rich intrusive rocks appear to have been formed mainly after the deposition of the Kinojevis and Blake River Groups to the north of the Kirkland Lake - Larder Lake Fault Zone, and were probably associated with the continuing development of the Round Lake Batholith. The time required to form the Round Lake Batholith is unknown, but it probably started with the partial melting of calc-alkalic volcanic rock during the deposition of the Pacaud Tuffs and continued during the deposition and subsidence of the Wabewawa, Catherine, and Skead Groups, on its eastern side. It probably continued to develop during the deposition and subsidence of the Lower Formation of the Tisdale Group, and Kinojevis Group on its northwestern side, where quartz monzonite and syenite now cut the batholith. The sedimentary rocks north of the batholith near the Kirkland Lake area, however, were probably not as deeply buried as elsewhere and were able to resist subsidence until isostatic balance was established in most areas surrounding the batholith. At this point, the sedimentary rocks were pulled down and southward to replace the subsiding eclogites and eclogitic residues formed by earlier partial melting. As the sedimentary rocks subsided and were transformed to eclogite facies rocks, magmas from the partial melting of these rocks rose to the surface to form the K-rich intrusive rocks and Timiskaming volcanic rocks. Sinking of the sedimentary rocks probably caused isoclinal folding in the overlying Larder Lake Group rocks.

GEOLOGICAL SIGNIFICANCE OF THE K–RICH IGNEOUS EXTRUSIVE AND INTRUSIVE ROCKS

If a final phase of K-rich felsic volcanism occurred at the end of each volcanic cycle, trachytes should have been recognized in the Hunter Mine Group by the author, and, in the Upper Formation of the Deloro Group by Pyke (1980). In addition, trachytic lavas should be more abundant in the Noranda part of the Abitibi Belt where the Blake River Group showed a more advanced stage of development than in the Kirkland Lake part. Instead, absence of trachytic lavas and tuffs as well as syenite, monzonite, and granodiorite intrusions are absent in the Noranda area. Therefore, it is doubtful if K-rich felsic volcanism is an integral part of the development of a volcanic cycle.

In the previous section on the Na-rich granitic batholiths, it was pointed out that although these batholiths cut the basal rocks of the Upper Supergroup, their parental rocks were probably the calc-alkalic rock of the lower supergroups. Similarly, the alkalic volcanic and intrusive rocks probably had parental rocks which belonged to the lower supergroups. In this case they should be considered part of the final phase in the transition of simatic to sialic crust connected with the lower supergroups. In the Kirkland Lake area, the alkalic volcanic rocks and K-rich intrusive rocks together show a spacial distribution similar to that of the sedimentary rocks thought to have been deposited off the edges of the earlier formed volcanic piles (*see* Figure 83).

The transformation of sedimentary rocks into K-rich igneous rocks by burial and partial melting would remove the last remaining vestiges of a calc-alkalic volcanic pile which was parental to a batholith. The transformation of sedimentary rocks flanking a calc-alkalic pile to K-rich igneous rocks probably would not occur simultaneously for all the sedimentary rocks. Instead, the transformation could occur sporadically as observed in the vicinity of the Round Lake Batholith and result in granitic rocks of different ages cutting all types of Archean rock. However, these rocks would tend to form the late phases of igneous activity in Archean terrains because they would generally require that their parental sedimentary rocks be buried by additional volcanic and sedimentary rocks.

Genesis of the Archean Crust

PRESENT–DAY ARCHEAN SHIELDS

The Archean crust beneath stable continental regions averages 40 km in thickness (Condie 1973; Ringwood 1975), is usually divided into upper and lower segments, and has seismic P-wave velocities of 5.8 to 6.4 km/second and 6.5 to 7.2 km/second, respectively (Figure 92). The seismic velocities tend to increase with depth, and in some regions, a discontinuity ("the Conrad") occurs between the upper and the lower crust. At the base of the crust, the P-wave velocity increases to about 8 km/second, at the Mohorovicic discontinuity.

Seismic P-wave velocities in the upper crust correspond to those of felsic igneous and metamorphic rocks similar in composition to granodiorite and quartz diorite, whereas the velocities in the lower crust are similar to those of basalts, diabase, and gabbro (Ringwood 1975). Because the lower regions of the crust have been subjected to temperatures from 300 to 600°C, mafic rocks would be highly metastable and would probably have been transformed into granulite and eclogite mineral assemblages with seismic P-wave velocity ranges of 7.3 to 8.4 km/second (Green and Ringwood 1967b). In trying to duplicate the P-wave velocities found in the lower crust, T.H. Green (1969) concluded that it must be similar in composition to gabbroic anorthosite in a high grade metamorphic facies (plagioclase + orthoclase + garnet + pyroxene ± kyanite); likewise, Ringwood (1975) suggested that the lower crust was equivalent to quartz diorite in the eclogite facies. Previously, den Tex (1965) had concluded that the lower crust was a heterogeneous mixtûre of 55% felsic and intermediate granulites and charnockites, 40% basic granulites and eclogites, and 5% ultrabasic rocks. Judging from the above studies, it seems likely that at present, rocks of felsic composition extend to the Moho. The question arises as to whether this felsic crust was formed during the Archean period, or subsequently.

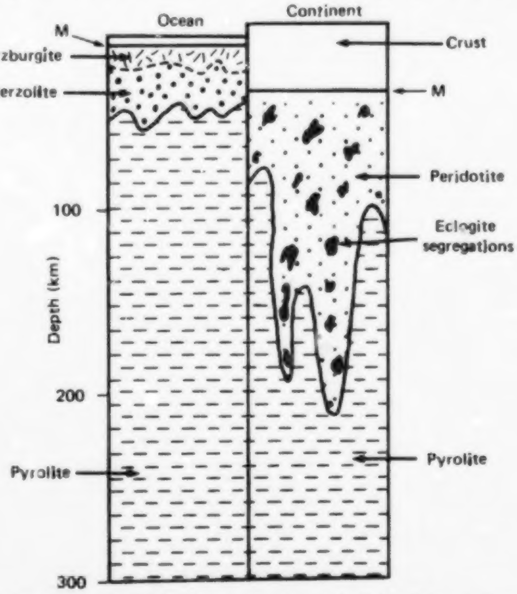


Figure 92. Chemically zoned model for the upper mantle of Ringwood (1979, p.10, Figure 3). Hartzburgite and Iherzolite are varieties of peridotite.

Proponents of a thin Archean crust have argued that geothermal gradients then were too high to form an extensive 40 km thick layer of granitic rock (Shacketon 1973a, 1973b; Fyfe 1973a, 1973b, 1973c, 1978a, 1978b; Hargraves 1976; O'Hara 1977; Tarney and Windley 1977). Many of these authors have suggested that the thickening of the Archean crust occurred either by underplating or by underthrusting during post-Archean events. Condie (1973), Wells (1976, 1979), and Burke *et al.* (1976) have all pointed out that mineral assemblages in some deeply exposed Archean terrains are indicative of geothermal gradients similar to those in the present day crust, and, in places, were even as low as 15°C/km. Based on this information, Condie (1973) and Wells (1979) concluded that the Archean crust in some places must have been at least 60 to 80 km thick. As well, O'Hara's (1977) observation that the Scourie gneisses were formed at a 50 km depth would indicate that granitic rocks during the Archean could exist at a level equivalent to that of the present mantle.

Further support for a thick Archean crust comes from isotope and REE data which indicate that magmas for Archean granitic rocks were generated from deeply buried parental calc-alkalic rocks, at depths of more than 100 km (Hanson and Goldrich 1972; Spooner and Fairbairn 1970; Moorbath 1975a, 1975b).

The Archean crust at the present time appears to be about 40 km thick, regardless of its present erosional level, and, therefore, the 40 km thickness must represent the approximate isostatic balance between the Earth's mantle and crust attained during the Archean and maintained since then (Wise 1974). Studies of the metamorphism in the Abitibi Belt show that less than 3 km of rock has been eroded since its formation (Jolly 1978); in contrast, in the gneissic terrains of Greenland and Scotland more than 40 km of rock have been removed (Wells 1976, 1979; O'Hara 1977). The implication from these studies is that different parts of the Archean crust were generated with different thicknesses. Since greenstone belts tend to be the youngest rocks with the lowest grades of metamorphism in most Archean terrains, it can be suggested that they represent thinner segments of the Archaean crust than the earlier gneissic crustal segments. Thus, it is concluded that some parts of the Archean crust have lost considerable thickness to erosion and others relatively little.

PRESENT–DAY SUBCONTINENTAL MANTLE

Ringwood (1975, 1979) stated that most of the present upper mantle is composed of pyrolite, a nonspecific pyroxene-olivine rock capable of yielding basaltic magmas on partial melting (Figure 93). Excluded from this description was the uppermost part of the mantle, considered to be mainly composed of peridotite (olivine-pyroxene) and eclogite (pyroxene-garnet) which have been strongly depleted in low-temperature melting components so that they do not produce basaltic magmas if partly melted. Much of

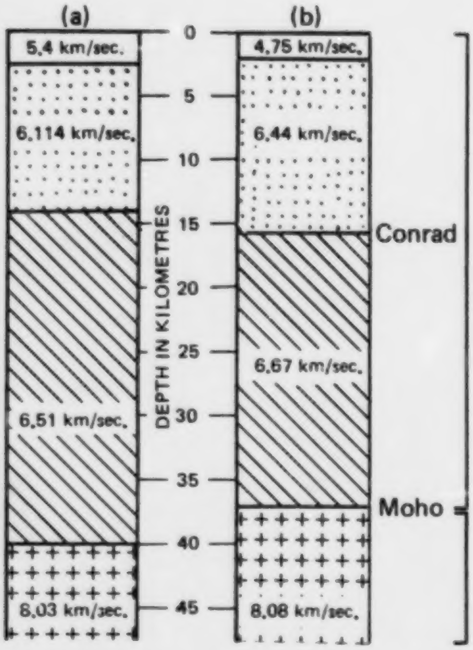


Figure 93. Crustal sections for 2 typical stable regions; the Precambrian shields in (a) central Wisconsin and (b) upper Michigan (from Steinhart and Meyer 1961).

the evidence for this theory comes from xenoliths found in alkalic basalts and kimberlites. Ringwood (1975) maintained that 2 groups of xenoliths occur, the peridotitic type with less than 15% garnet, and those of eclogitic association with more than 30% garnet found only in kimberlite pipes in stable continental crust. He further suggested that these eclogite xenoliths were formed from partly melted quartz eclogite which was originally oceanic tholeiite transferred to mantle depths. He noted that quartz eclogite was rare in comparison to other types of eclogites.

Of the garnet peridotitic xenoliths which constitute the largest proportion of the xenoliths in kimberlite (Green 1975), many were found to have retained their low melting components by O'Hara *et al.* (1975). This observation, plus the ^{87}Sr/^{86}Sr ratios led Brooks *et al.* (1976) to conclude that the age of the uppermost mantle below the continental crust was 300 Ma, or more, and undepleted. This conclusion conflicts partly with the studies of Ringwood (1975, 1979). Instead, the uppermost mantle is probably composed of both depleted and undepleted garnet peridotite; the peridotite as well as the eclogite are of several types.

A lithosphere up to 300 km thick may exist below the continental crust (Brooks *et al.* 1976; Davies 1979). Much of the lithosphere might consist of refractory peridotite (as suggested by Ringwood 1979), being the residuum of the partial melting of the Archean mantle during the production of komatiitic and tholeiitic magmas. If the primary crust on which the komatiitic and tholeiitic lavas were deposited was carbonaceous chondrite which had escaped partial melting, a large proportion of the uppermost mantle could still be undepleted peridotite and garnet peridotite older than 300 Ma. These depleted and undepleted peridotites would have incorporated eclogite during its formation and sinking, concurrent with the production of calc-alkalic volcanic and granitic rocks in the Archean.

PRIMITIVE CRUST

The oldest Archean rocks for which records exist formed at the surface of the earth either as volcanic rocks or as sedimentary rocks. In the map area for example, the Stoughton - Roquemaure Group seems to have been deposited on a surface (primary crust), or on a thin veneer of sedimentary rock overlying a primary crust between the volcanic piles represented by the lower supergroups. Likewise, the initial komatiitic and tholeiitic lavas of the older volcanic cycles which culminated in the Hunter Mine Group and Pacaud Tuffs must have been deposited on a base, possibly a primary crust or a thin sequence of sedimentary rocks derived from even earlier volcanic piles.

Previous hypotheses concerning Archean crustal development have supported either a mafic or a felsic primary crust. Proponents of a mafic crust (Anhaeusser 1971a, 1971b; Glikson 1972; Moorbath 1976; Hanson 1978) consider it to have been an oceanic one, composed of komatiitic and tholeiitic lavas formed either along rift zones or at spreading oceanic ridges. In the Kirkland Lake area, evidently calc-alkalic volcanic and sedimentary rocks were succeeded by komatiitic and tholeiitic lavas which, therefore, could not have been unique to the primary crust.

The hypotheses postulating the existence of a felsic to intermediate primary crust present it as being thin, widespread, and mobile (Ramberg 1964; Bridgwater and Fyfe 1974; Hargraves 1976; Shaw 1976; Collerson and Fryer 1978; and Gorman *et al.* 1978). The thin sialic crust is considered to be the result of a molten stage occurring shortly after the Earth's formation because of the heat liberated from accretion, core formation, and radioactivity (Fanale 1972; Shaw 1976). During subsequent cooling, according to Hargraves (1976) and Fyfe (1976), felsic rocks would likely have formed at the surface, followed by the crystallization of more mafic rocks at depth underplating the felsic crust. Consequently, the Earth's geothermal gradients are thought to have been 2 to 4 times steeper, with the crust much thinner during the Archean than at present (Shaw 1976; Lambert 1976; Fyfe 1973, 1976; Bickel 1978; O'Nions and Pankhurst 1978). This would mean that later formed, thick, mafic volcanic successions, would have been unsupportable at the surface (Gorman *et al.* 1978).

Others do not believe that the Earth's mantle underwent a molten stage. According to Grossman (1972), Clarke *et al.* (1972), Hutchinson (1974), and Murthy (1976), the Earth may have formed its core before accreting its mantle, thus avoiding the necessity of melting the mantle. The presence of rare gases ^{4}He and ^{90}Ar in today's mantle and atmosphere is inconsistent with a mantle that underwent a molten stage (Fisher 1976; Hart *et al.* 1979). Based on Rb-Sr isotope ratios, Jahn and Nyquist (1976) concluded

TABLE 9: THE KIRKLAND LAKE MEGACAULDRONS OF THE ABITIBI BELT.

Stage 1: Extrusion of komatiitic and tholeiitic lavas from the mantle.

Upper Supergroup: Stoughton - Rouquemaure, Larder Lake, Lower Formation of the Tisdale and Kinojevis Groups.

Lower Supergroups: Wabewawa and Catherine Groups and Lower Formation of Deloro Group. As well, komatiitic and tholeiitic lavas are assumed to have existed in basal parts of the volcanic piles represented by the Hunter Mine Group and Pacaud Tuffs.

Stage 2: Development of calc-alkalic volcanic rocks from the partial melting of "eclogized" komatiitic and tholeiitic volcanic rocks.

Upper Supergroup: Blake River Group.

Lower Supergroups: Skead and Hunter Mine Groups and Pacaud Tuffs.

Stage 3: Generation of granitic rocks from the partial melting of "eclogized" calc-alkalic volcanic and sedimentary rocks.

Upper Supergroup: Flavrian Pluton.

Lower Supergroups: Lake Abitibi, Round Lake, and Watabeag Batholiths, and many granitic stocks and dikes.

that the Archean mantle was not homogenized by mantle convection, nor were the mantle sources for Archean basaltic magmas depleted in REE by earlier melting according to Sun and Nesbitt (1977, 1978a) who suggested that the mantle was similar in composition to undepleted carbonaceous chondrite. Other implications derived from these studies are that igneous rocks were limited prior to the earliest formed Archean rocks, and that the initial komatiitic and tholeiitic lavas could have been extruded onto a surface of carbonaceous chondritic material, as maintained by Dietz (1965). He argued further that a magnitude of a billion years in time would likely be required before sufficient heat from radioactive sources could build up and initiate mantle convection.

Therefore, the surface on which the Abitibi Belt developed was likely to have been composed of carbonaceous chondritic material formed about 4500 to 4600 Ma ago. At mantle depths, the carbonaceous chondrite graded into peridotite and garnet peridotite of the same chemical composition. Also, a crust and mantle of carbonaceous chondritic composition would have provided a stable platform on which to accumulate thick Archean ultramafic to mafic volcanic successions, and ultimately was replaced by a thick granitic crust. Thus, the carbonaceous chondrite surface may be regarded as a primitive crust in that its physical properties changed with depth, even though the chemical composition was fairly constant during Archean times.

FORMATION OF THE ARCHEAN CONTINENTS BY MEGACAULDRONS

The presently exposed Archean crust in the Kirkland Lake part of the Abitibi Belt was formed by the repetition of a 3-stage sima-to-sial transformation (Table 9). In the first stage, large volumes of magma rose from the mantle to produce a thick crust composed of ultramafic to mafic volcanic and intrusive rock. During the second stage, parts of this crust sank and were partly melted resulting in mafic to felsic magmas that returned to the surface and there accumulated as calc-alkalic volcanic piles surrounded by sedimentary successions. In the third stage, these calc-alkalic volcanic and sedimentary rocks subsided to deep crustal and mantle depths, where, during partial melting, they created felsic magmas which produced felsic stocks and batholiths. During the 3 stages, the dominant displacement of surface material occurred by downward faulting, buckling, and tilting of crustal rocks concomitant with the upward movement of magma replenishing the surface rocks.

Because this process is grossly similar to the processes of volcanic eruption and subsidence of the overlying material into the magma chambers, these areas of sial-to-sima transformation are termed megacauldrons. Such megacauldrons represent both the locations and the processes by which the primary crust is converted into the continental crust of the Archean shields.

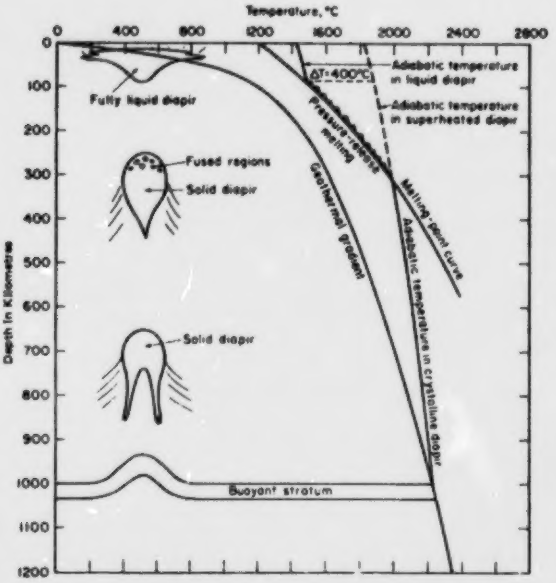


Figure 94. Magma generation model of Ramberg (1972, p.58, Figure 10) based on perturbation in buoyant stratum, diapiric rise of a solid mass, and pressure-release melting.

INITIATION OF DIAPIRS

In this model, the primary crust is assumed to have been an accreted heterogeneous mixture of meteoritic material consisting mainly of carbonaceous chondrite which at mantle depths is expected to have graded into a dense rock resembling peridotite or pyrolite. Magmas were created as the buildup of heat in the peridotite initiated the rise of diapirs (Figure 94). According to Ramberg (1972), a thermally isolated mass rises as a solid until pressure is sufficiently reduced for melting to begin at a depth of about 300 km; also, the adiabatic decompression is thought to be adequate to achieve complete melting when the mass reaches a depth of about 80 km. The Archean diapirs were possibly very large in volume and area, and may have been the principal mechanism of upward transfer of heat during this period (Bickle 1978).

THE PRIMARY MEGACAULDRON

The primary megacauldron (Figure 95) is thought to progress in the following manner:

Stage 1 Development of a Mafic Crust

Once the magma from a diapir has broken through the surface, mafic and ultramafic komatiite volcanism produces large volumes of lava which accumulate in the region above the diapir. Subsidence of the lavas by a combination of downfaulting and sagging of the primary crust occurs above the diapir (Figure 95A). As higher satellitic magma chambers develop in the central area of subsidence, tholeiitic volcanism replaces komatiitic volcanism as a result of the fractionation of olivine, pyroxene, and plagioclase which remain at depth as dunites, pyroxenites, and gabbros (Figure 95B). The thickness of ultramafic and mafic rock above the diapir is proportional to the size of the diapir and largely contained above it by the subsidence.

Stage 2: Development of a Calc-alkalic Rock

The accumulation of the komatiitic and tholeiitic lavas is accompanied by the conversion of the sinking ultramafic and mafic rocks to eclogites which in turn undergo partial melting to produce calc-alkalic magmas which come to surface to build a calc-alkalic volcanic pile (Figure 95C). The remaining garnetiferous residua continue sinking.

As the calc-alkalic volcanic pile accumulates, its core subsides because the eclogite continues to sink. As well, aprons of clastic and chemical sedimentary rocks are deposited on the edges of the calc-alkalic pile largely outside the area of subsidence. During this process, the ultramafic and mafic rocks of the earlier komatiitic and tholeiitic volcanism are nearly all consumed.

Stage 3: Development of a Granitic Crust

As the calc-alkalic volcanic rocks subside, they are transformed into garnet-bearing amphibolites and eclogites and continue to sink to mantle depths (Figure 95D). These eclogites melt partly to produce intermediate to felsic magmas strongly depleted of HREE. The resulting garnet-rich residua sink, and the magmas, on migrating upward, allow plagioclase to fractionate so that they now become trondhjemite. Meanwhile, granulites are developed at depth in the volcanic pile.

The generation of K-rich magmas are considered part of the third stage. This is because their sources are likely to have been sedimentary rocks deposited peripherally to the calc-alkalic pile. Before the sedimentary rocks can produce K-rich magmas, these sedimentary rocks first require burial by volcanic and sedimentary rocks generated from subsequent diapirs coming to surface near the primary diapir.

The megacauldron caused a trondhjemitic batholith to intrude felsic calc-alkalic volcanic rocks that graded into clastic and chemical sedimentary rocks away from the batholith. Komatiitic and tholeiitic lavas would be notably absent.

SUCCEEDING MEGACAULDRONS

Additional diapirs may rise toward the Earth's surface near the initial diapir, possibly within a few million years after its creation, to a few hundred million years later. Komatiitic magmas generated from a diapir directly below the initial one, would not likely penetrate the thick granitic column (Fyfe 1978a, 1978b). Instead, the magmas would be deflected and come to the surface beside the granitic column and form an "adjacent" megacauldron. Magmas of a diapir 20 to 50 km or more away from the initial diapir would not suffer interference while rising. If the diapir is near enough for its surface phenomena to overlap a primary megacauldron, it is a proximal megacauldron.

In Figure 96, examples of subsequent megacauldrons are displayed; the left one, the adjacent megacauldron (b) results in an enlargement of the granitic terrain, and the right one, the proximal megacauldron (c), culminates in a greenstone belt between the 2 megacauldrons. At this point in the

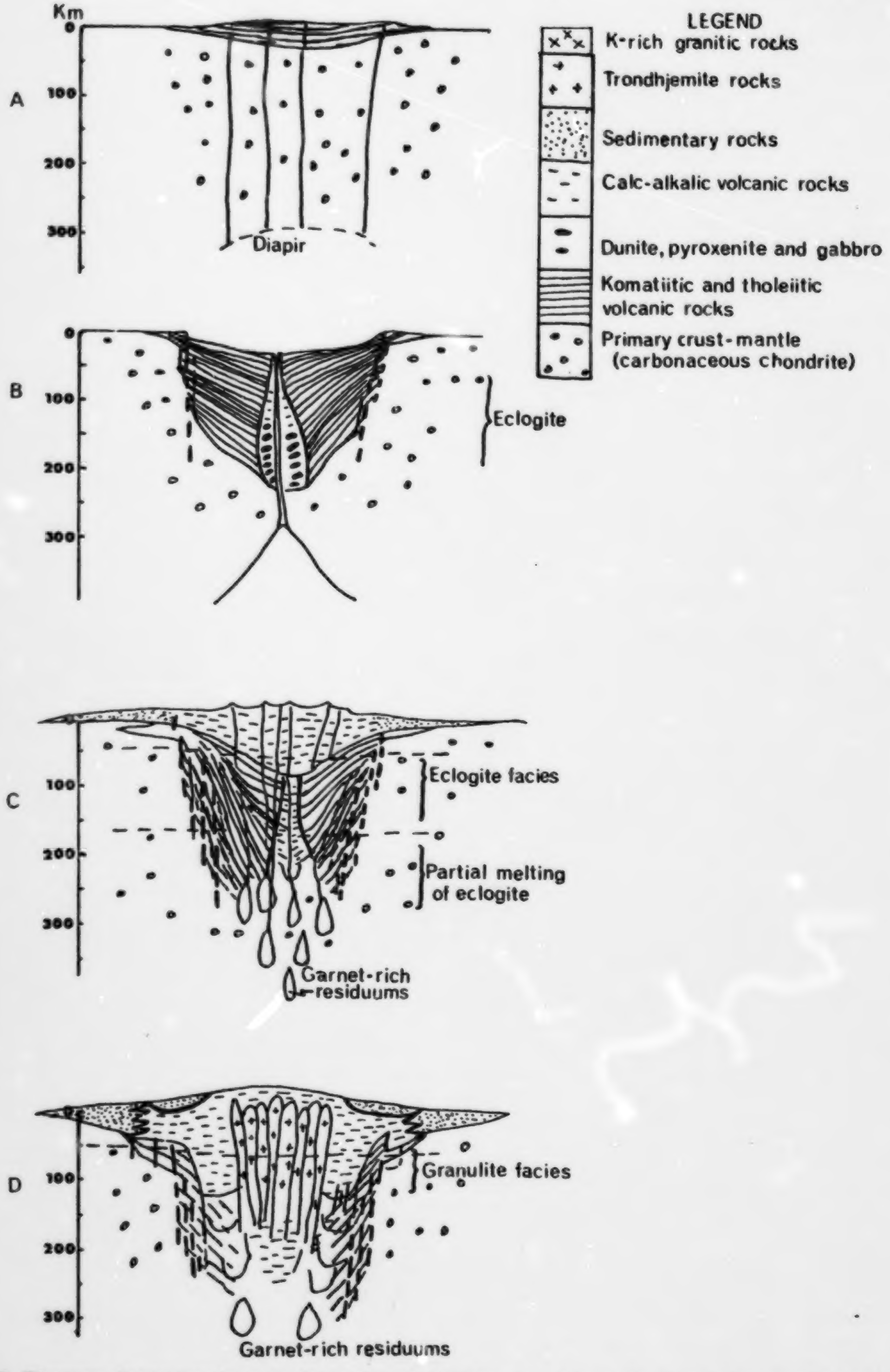

Km
0
100
200
300
A
Diapir
B
Eclogite
C
Eclogite facies
Partial melting of eclogite
Garnet-rich residuums
D
Granulite facies
Garnet-rich residuums
LEGEND
K-rich granitic rocks
Trondhjemite rocks
Sedimentary rocks
Calc-alkalic volcanic rocks
Dunite, pyroxenite and gabbro
Komatiitic and tholeiitic volcanic rocks
Primary crust-mantle (carbonaceous chondrite)

Figure 95. Diagram displaying the transformation of simatic crust into sialic in a primary megacauldron: A. the eruption of komatiitic magmas from a rising mantle diapir; B. the sinking of the volcanic rocks and the eruption of tholeiitic magmas; C. the partial melting of eclogized metavolcanics to produce calc-alkalic magmas; D. the partial melting of eclogized calc-alkalic volcanic rocks to form trondhjemitic magmas.

development of the subsequent megacauldrons, the primary megacauldron may be still in the process of forming a batholith, and the tectonic stresses induced by the developing adjacent megacauldron produce regional anisotropic fabrics in the granitic rocks.

Adjacent Megacauldrons

The left megacauldron in Figure 96 rises beneath the margin of the primary megacauldron and differentiates along side it as has been described for the primary one. The initial komatiitic lavas had been deposited on sedimentary rocks. Felsic tuffs of the earlier calc-alkalic volcanic pile, cause the sedimentary rocks to become deeply buried and transformed into eclogite. This process is complemented by the sinking of eclogite in the primary megacauldron, thus drawing the sedimentary rocks converted to eclogite down and inward, towards the base of the batholith. These sedimentary rocks converted to ecologite are then partly melted to produce K-rich magmas which cut the trondhjemitic rocks. Only remnants of greenstone might remain.

Proximal Megacauldrons

The megacauldron, c, on the right-hand side of Figure 96 proceeds independently of the first one, a. The komatiitic and tholeiitic lavas deposited on the sedimentary and felsic calc-alkalic rocks are not involved in the subsidence at the core of the developing ultramafic to mafic volcanic successions. Instead, these lavas subside, and form a trough between the 2 megacauldrons, a result of the sinking eclogite in the primary megacauldron. As a calc-alkalic pile grows and expands in the younger megacauldron, calc-alkalic tuffs and sedimentary rocks are deposited, disconformably or unconformably, on the komatiitic and tholeiitic lavas in the trough.

At the base of the trough, the earlier (deeply buried) sedimentary rocks are converted to eclogite, and on sinking form K-rich granitic magmas which migrate upward to cut the overlying volcanic and sedimentary rocks. Meanwhile, denudation of the granitic rocks belonging to the primary megacauldron allows arkosic sediments to be deposited in the centre of the trough.

The volcanic and sedimentary rocks in the trough would contain many features found in most Archean greenstone belts. For instance, these rocks would appear as supracrustal rocks in a *milieu* of granitic rock (Goodwin 1968; Anhaeusser *et al.* 1969; Glikson 1972; Talbot 1973; Langford and Morin 1976; Moorbath 1976; Glikson and Lambert 1976); and as well, they would show low-grade metamorphism in contrast to the high-grade metamorphism of the adjoining granitic terrains (Anhaeusser *et al.* 1969). Granitic rocks could also form the basement for some komatiites (Baragar and McGlynn 1976; Nisbet *et al.* 1977) but in general, the volcanic rocks would tend to be a bimodal assemblage of mafic and ultramafic lavas interlayered with felsic calc-alkalic tuffs, as observed by Glikson (1976), Glikson and Lambert (1976), Barker and Peterman (1974), Shaw (1976), and Hubregtse (1976). Sedimentary rocks would predominate at the top of the succession contained in the trough, as in the Barberton Greenstone Belt (Anhaeusser 1973), in the Rhodesian greenstone belts (Hawkesworth *et al.* 1975) and in the Australian greenstone belts (Glikson 1976a).

The dips of the rocks in the trough would be steep (Anhaeusser *et al.* 1969) and in places, especially where close to the batholiths, the rocks could be overturned, particularly in the most deeply eroded terrains, and thus be misinterpreted as thrust napps (Stowe 1974; Coward and James 1974; Myers 1976; Talbot 1979; Gorman *et al.* 1978; Thurston and Breaks 1978; and Blackburn 1980). In the Kirkland Lake area, the volcanic rocks are overturned on the southern side of the Round Lake Batholith (Moorhouse 1941).

OTHER MODELS OF ARCHEAN CRUSTAL DEVELOPMENT

The various models proposed to date for greenstone belts in Archean crustal terrains, some of which have been reviewed by Windley (1977, p.45-64), fall into 2 main groups:

1. those incorporating the principles of plate tectonics
2. those relying mainly on vertical tectonic movement in a restricted area.

Examples of plate tectonic models are those of Goodwin and Ridler (1970), Rutland (1973), Tarney *et al.* (1976), Langford and Morin (1976), Windley (1977a, 1977b, p.63), and Blackburn (1980). Vertical tectonic models have been presented by Anhaeusser *et al.* (1969), Anhaeusser (1971), Glikson (1972, 1976a, 1976b, 1976c, 1976d), Condie and Hunter (1976), Baragar and McGynn (1976), Gorman *et al.* (1978), and Collerson and Fryer (1978).

HORIZONTAL TECTONIC MODELS

Difficulties arise in attempting to apply plate tectonic concepts to Archean greenstone belts (Glikson 1976b; Hargraves 1976; Shaw 1976; Lambert 1976; Baer 1977; Sun and Nesbitt 1978a, 1978b; Taylor 1979; and Kroner 1977). Based on the proceeding discussions, however, it can be pointed out that if the Archean crust had developed with thicknesses as great as those of the present crust, it should have been competent enough to support and preserve ophiolitic successions indicative of Benioff zones. To the author's knowledge, no ophiolites have been recognized in Archean terrains. If Benioff zones existed, the calc-alkalic volcanic and granitic rocks should show isotope and REE patterns similar to those of Cenozoic island arcs. Instead, such patterns in Archean calc-alkalic volcanic and granitic rocks are compatible with generation directly from the partial melting of eclogite and garnet-bearing amphibolite whereas those in Cenozoic rocks are not.

Existing plate tectonic models require sialic plates which could readily separate and collide with each other.

Kirkland Lake Area

In an attempt to devise a plate tectonic model for the Abitibi Belt, the model after Tarney *et al.* (1976; Figure 97) can be used. It seems logical to suggest that

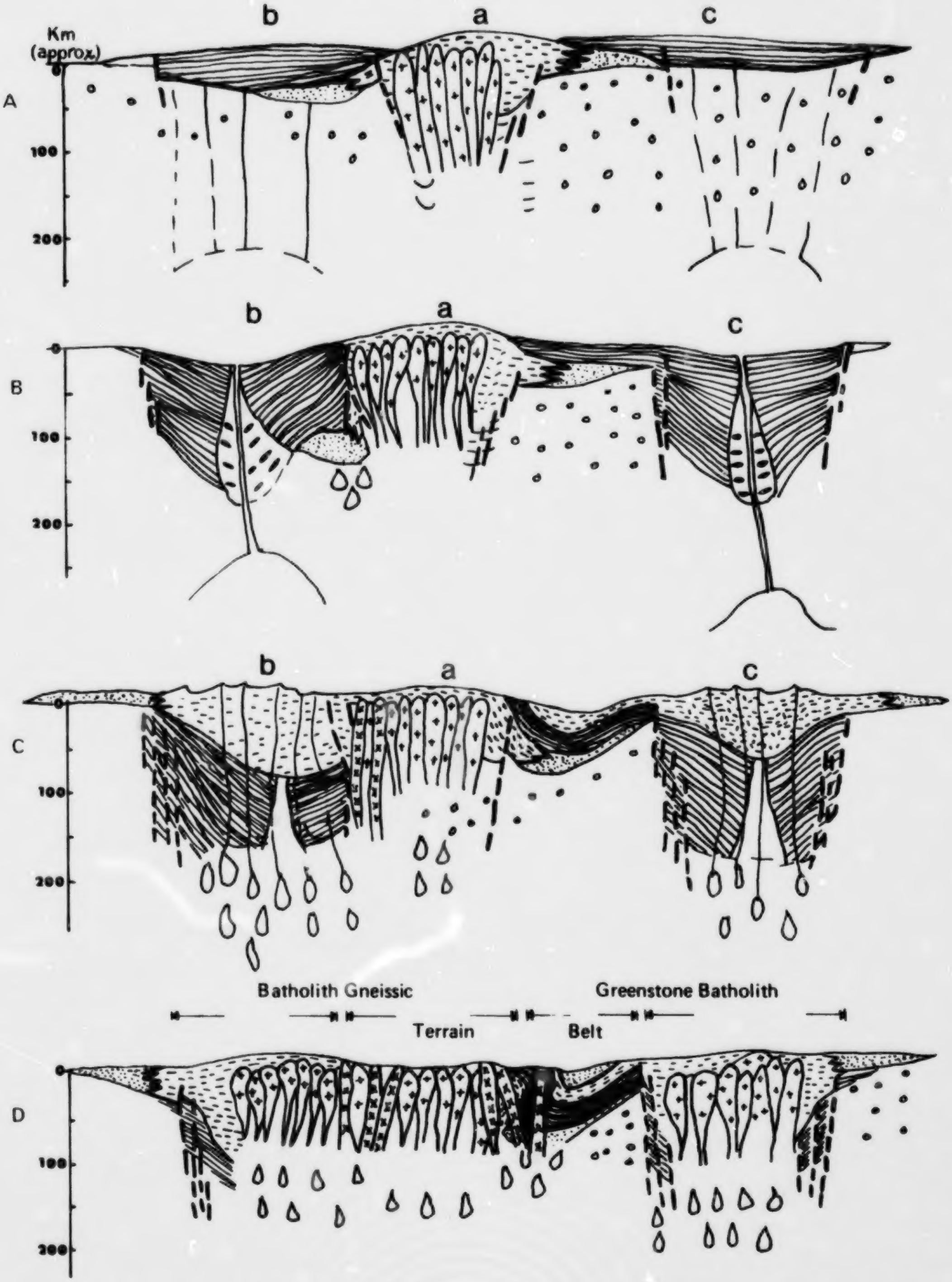


Figure 96. Model of the evolution of Archean shields (*see* legend of Figure 95). Areas a, b, and c represent primary, subjacent, and adjacent megacauldrons, respectively.

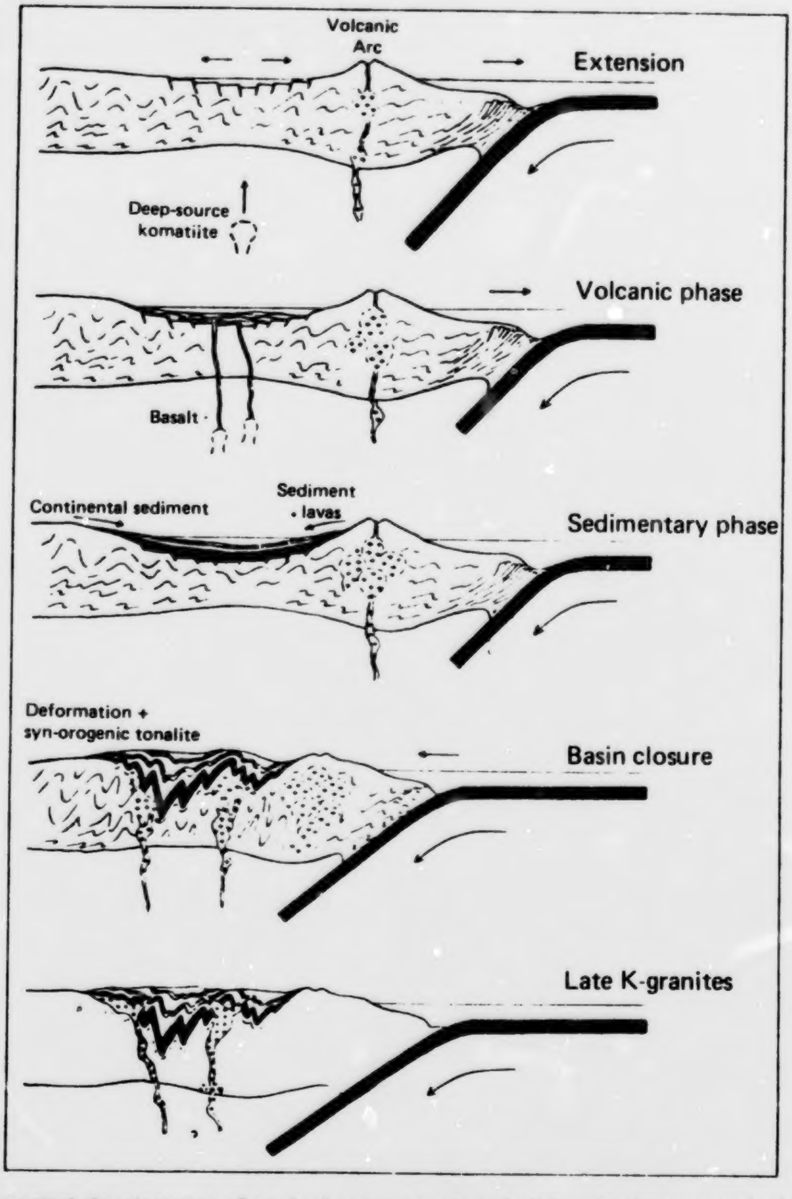


Figure 97. Suggested development of an Archean greenstone belt in a back-arc marginal basin position on comparison with the "Rocas Verdes" complex in S Chile. Back-arc extension in an Archean ductile lithosphere produces a greater component of crustal thinning rather than crustal rifting, as today. Magma production exceeds that required for simple extension. Sediment source is mixed volcanogenic/continental and the sequence may include calc-alkalic andesitic and salic lavas from the adjacent volcanic arc. Later movement of arc towards continent produces deformation and synclinal form of greenstone belt. Andean-type tonalitic - to - granitic plutons (derived from mantle by at least 2-stage fractionation process) may be syn- to post-tectonic with compositions dependent on depth of melting of subducted oceanic crust (after Tarney, Dalziel, and de Wit 1976).

the volcanic arc might have been the Hunter Mine Group, cut by the Lake Abitibi Batholith, and that the continental side was likely the Round Lake Batholith (as proposed by Goodwin and Ridler 1970, and Ridler 1975) with the volcanic rocks in between filling a back-arc marginal basin. But this comparison poses a problem, since no extension appears to have occurred south of Timmins in the Upper Formation of the Deloro Group (*see* Figure 5), which predates the Hunter Mine and Skead Groups in the Kirkland Lake area. Thus, any extensional tectonics were probably more local than required for plate tectonics and were caused either by diapiric doming or cauldron subsidence. The back-arc basin was filled with komatiitic and tholeiitic lavas, but without continental sedimentary material. A volcanic island-arc pile (Blake River Group) would not be expected to have formed in the core of the basin unless calc-alkalic volcanism suddenly shifted to the centre of the basin which is opposite to the usual tendency of calc-alkalic volcanism to migrate in the fore-arc direction. Another problem would be to account for the lack of tectonic deformation which should have developed in the rocks during the basin closure, since fault contacts are absent between the groups of volcanic rock; there is no evidence of major horizontal movements along the Destor - Porcupine and Kirkland Lake - Larder Lake Fault Zones.

VERTICAL TECTONIC MODELS

Most vertical tectonic models involve downwarping through the folding or faulting of a pre-existing oceanic or continental crust (for example, Figure 98). Those dealing with an earlier continental crust suggest that it must have been buried by mafic volcanic rocks, calc-alkalic rocks, and sedimentary rocks, filling a subsiding trough so that the continental crust became deformed and partially melted, thus producing felsic magmas which, on returning to surface cut the supracrustal rocks. Critics of this model point out that the isotopic and trace element studies on Archean granitic rocks indicate that these rocks are not remelted granitic rock, but, instead are newly generated additions to the earth's crust (Moorbath 1975a, 1975b; Glikson 1979b). In addition, xenoliths of greenstone occur in the oldest known Archean granitic rocks, showing that there were mafic volcanic rocks as old or older than the granitic rocks.

Proponents of a pre-existing oceanic crust (for example, Glikson 1972, 1976a, 1976b, 1976c, 1976d, 1979a, 1979b, 1979c) suggest that partial melting of an oceanic crust at its base gave rise to the granitic rocks found in Archean terrains as shown in Figure 98. In his criticism of this model, Windley (1977) pointed out that mafic remnants of greenstone belts in gneissic terrains cannot be equivalent to the lower ultramafic - mafic parts of later, better preserved greenstone belts, and that in places, komatiitic lavas unconformably overlie granitic rocks (Bickle *et al.* 1975).

An additional criticism of Glikson's model is that unless the primary oceanic crust was about 400 km thick, it would be difficult to generate the volume of granitic rock presently observed in Archean granitic terrains. Although Glikson's model does not adequately explain the generation of granitic rocks, nor allow for the development of thick calc-alkalic successions such as the Blake River Group, it does provide a model free of the problems of applying plate tectonic concepts to Archean terrains.

APPLICATION OF THE MODEL

Relative to other greenstone belts, the Abitibi Belt tends to have a higher ratio of volcanic to granitic rock. Volcanic successions comparable to the Blake River and Kinojevis Groups are rarely as well preserved in granitic greenstone terrains elsewhere. Thus, with lesser amounts of these 2 groups of rock and with additional granitic rock, the Abitibi Belt could show a closer resemblance to other belts. Therefore, before the model is used to interpret the geological development of the Archean crust elsewhere, a few conjectures concerning the evolution of the Abitibi Belt may be worthwhile.

If the Kinojevis and Stoughton - Roquemaure Groups had been thicker and had subsided deeper between the Destor - Porcupine and the Kirkland Lake - Larder Lake Fault Zones, the Blake River Group could have been more extensively developed; except that the Flavrian Pluton likely would have become a batholith, thus obliterating much of the Blake River Group and occupying much of the area between the 2 fault zones (*see* Figure 5). Then, there would be a close resemblance between this part of the Abitibi Belt, and, for example, the greenstone belts within the Wabigoon Subprovince of northwestern Ontario (Figure 99). In both cases, subvertical median fault zones occur, or would be expected to occur in the greenstone areas between the trondhjemitic batholiths.

Pursuing this idea further, if the Flavrian Pluton had expanded to replace all of the calc-alkalic rocks between the 2 fault zones, the remaining volcanic rocks would probably have been komatiitic, and tholeiitic lavas would have been interlayered with a few calc-alkalic tuffs and tuff-breccias belonging to the Stoughton - Roquemaure and Larder Lake Groups (*see* Figure 5). Some sedimentary rocks and calc-alkalic felsic tuffs deposited peripherally to the pre-existing calc-alkalic volcanic pile would likely overlie parts of the remaining komatiitic and tholeiitic volcanic successions as shown in Figure 98. In reality, several of the Archean greenstone belts in southern Africa and Australia are believed to be of this type (*see* Figure 96).

THE SUBPROVINCES OF NORTHWESTERN ONTARIO

The Uchi, Wabigoon, and Wawa Subprovinces shown in Figure 99 are composed of numerous low-grade metamorphic greenstone belts that enclose large trondhjemitic batholiths. These subprovinces alternate with the English River and Quetico Subprovinces formed mainly of granitic and sedimentary rocks, all of which, except for the K-rich granitic rocks, have undergone high-grade metamorphism (Breaks *et al.* 1978). Subvertical boundary faults occur in most places along the contacts between the subprovinces (Schwerdtner *et al.* 1978); each subprovince tends to have rocks of similar age along its strike length, the older rocks being in the English River and Quetico Subprovinces (Goodwin 1978; Blackburn 1980).

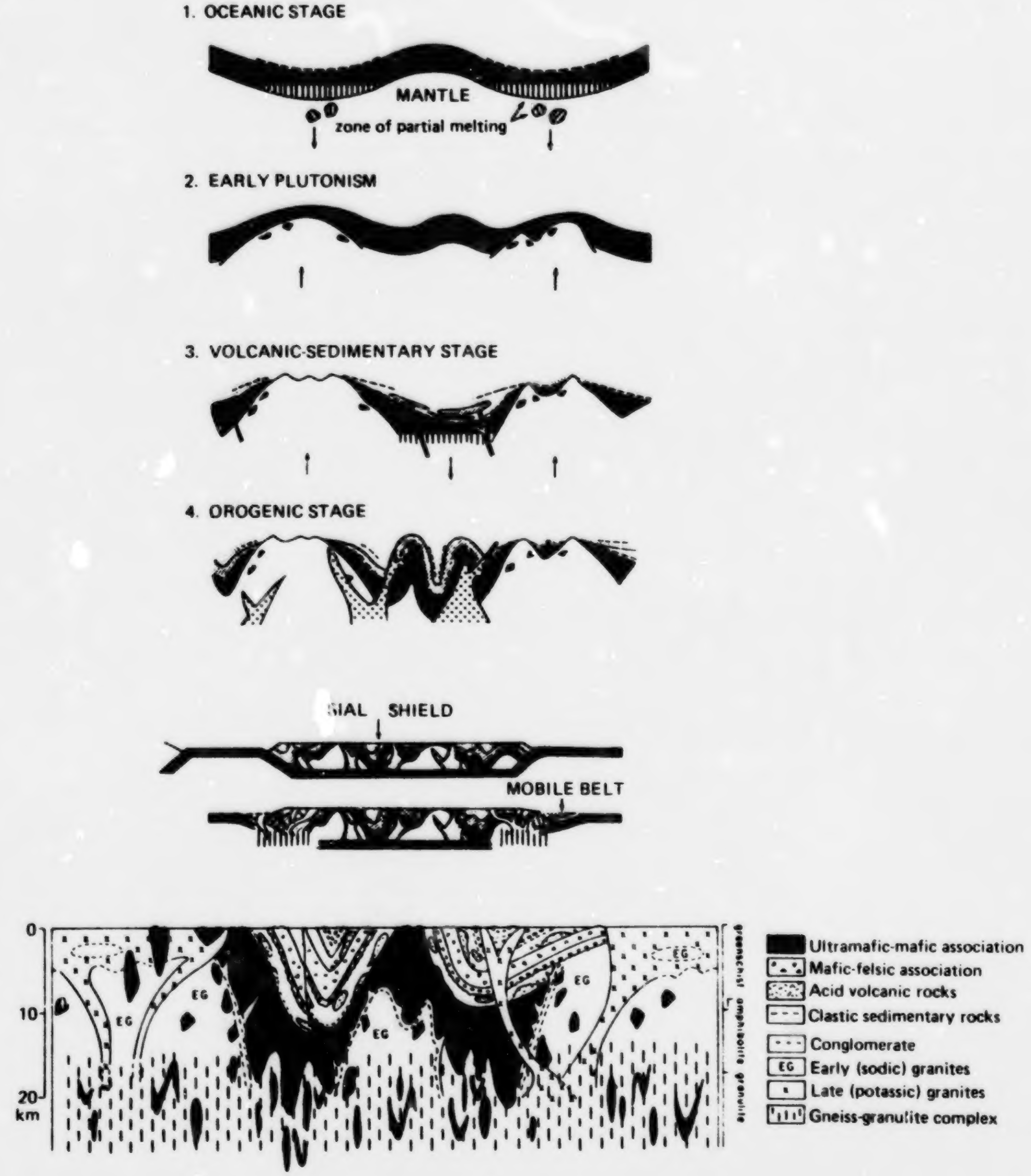
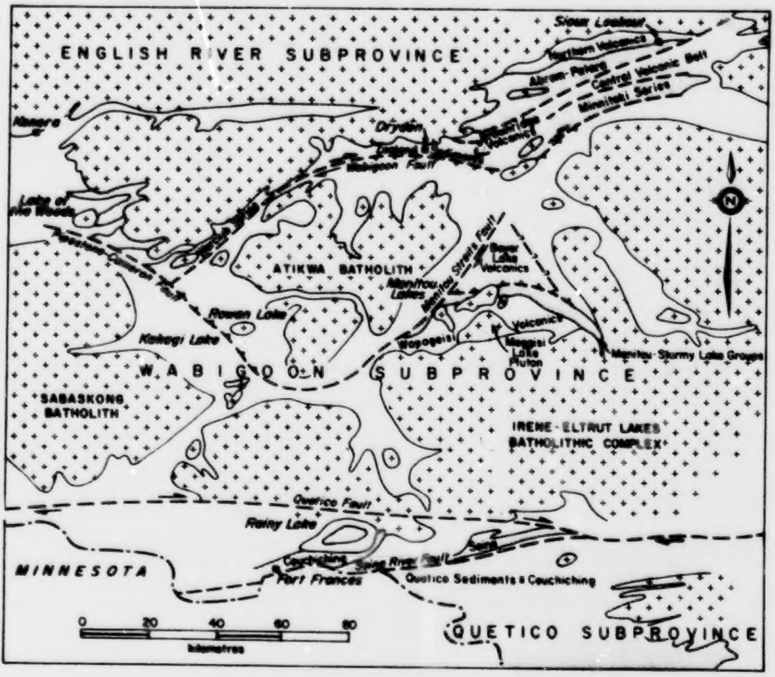


Figure 98. A model of evolution of Archean shields. 1: Oceanic stage; mega-rippling of the oceanic crust, minor sedimentation of chert, ferruginous chert, banded iron formations and pelitic sediments derived through erosion of the oceanic crust at structural highs. Partial melting of subsiding eclogite and/or amphibolite segments of the crust. 2: Early plutonism; low degrees of partial melting of the oceanic crust gives rise to sodic acid magmas which rise as diapiric oval-shaped or elongated batholiths. 3: Volcanic-sedimentary stage (greenstone belts); the isostatic rise of the early batholiths associated with the subsidence of intervening tracts of the ocean crust. Further partial melting of the oceanic crust below these troughs gives rise to calc-alkalic volcanism. The erosion of the batholiths as well as of the intrabasinal volcanics results in the accumulation of detrital sediments. 4: Orogenic stage; further subsidence of the volcanic-sedimentary troughs between the granitic nuclei into warmer regions of the crust results in folding and low-grade metamorphism. Low degrees of partial melting at the base of the thickening crust give rise to younger (potassic) granites. Shield formation stage cooling of the volcanic-sedimentary troughs results in the aggregation of the granitic nuclei into shields. The boundaries of these shields with the surrounding oceanic crust constitute the loci of development of high-grade "mobile belts", possibly above subduction zones (after Glikson 1972).

Figure B

Prominent lithologies and structural features of the western end of the Wabigoon and adjacent subprovinces in Ontario. Granitoid rocks (crosses) include felsic plutonic, subvolcanic, and gneissic terraines, and some mafic rocks. Sedimentary rocks (shaded) are clastic, with some chemical sediments. Volcanic rocks are shown without pattern. Faults (broken lines)



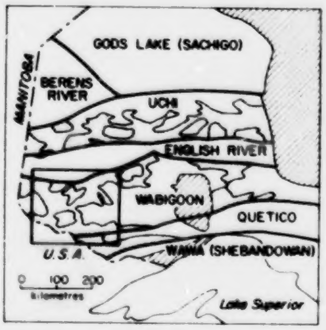


Figure A

Sketch map showing the subprovinces of the Archean Superior Province in northwestern Ontario, and location of Figure 2. Archean supracrustal rocks (shaded) are both volcanic and sediments; Proterozoic (striped) is cover.

Figure 99. General geology of the western part of the Wabigoon Subprovince (modified after Blackburn 1980).

In the English River Subprovince, Breaks *et al.* (1978) recognized a southern domain of gneissic plutonic rock, 3040 Ma old, cut by massive Na-rich granitic rock intruded in turn by K-rich granitic rock about 2600 to 2700 Ma old (Figure 100). A northern domain of metasedimentary rock was found cut by small intrusions of Na- and K-rich granitic rocks. The metasedimentary rocks contain clastic detritus which was thought to have eroded from coeval volcanic rocks in the Uchi Subprovince to the north and from an elevated granitic terrain positioned within the southern domain (Breaks *et al.* 1978).

Goodwin (1978a, 1978b) maintained that because of their low $^{87}Sr/^{86}Sr$ initial ratios and steep REE patterns, the 3040 Ma old gneisses of the English River Subprovince were generated from partly melted mafic crustal rocks. This would suggest that the proposed model is applicable to the rocks in the English River Subprovince as well as to the rocks in the Wabigoon Subprovince as indicated previously.

Following the model, the English River Subprovince probably represented a westerly trending row of diapirs, before *circa* 3000 Ma ago, which culminated in trondhjemitic batholiths intruding calc-alkalic volcanic piles with prisms of sedimentary rocks to the north and south (Figure 101). Succeeding this, rows of diapirs ascended toward the surface, north and south of the row of earlier diapirs. The diapirs of the Wabigoon Subprovince produced a succession of mafic and ultramafic lavas, that buried the sedimentary rocks, and caused them to sink downward and inward toward the base of the Na-granitic rocks of the English River Subprovince. Here the rocks were partly melted and resulted in K-rich granitic intrusions 2600 to 2700 Ma old, cutting the southern domain (Figure 101).

Because of their distance from the primary English River diapirs, the lavas of the Uchi diapirs did not deeply bury the northern prism of sedimentary rocks (Figure 101). Instead, as calc-alkalic piles were developed in the Uchi Subprovince, more sedimentary rocks were added to the succession of sedimentary rocks and only limited amounts of granitic rocks were produced along with the metamorphism of the sedimentary rocks (Figure 101). The fault boundaries between the subprovinces probably originated during the development of each sima-to-sial transformation and remained active until an isostatic balance between the mantle and the rocks of each subprovince was achieved. The diapirs responsible for the rocks of the English River Subprovince were probably larger and less numerous than those responsible for the rocks in the Uchi and Wabigoon Subprovinces.

BARBERTON GREENSTONE BELT, SOUTH AFRICA

Until recently, the Barberton Greenstone Belt and its surrounding granitic rocks (Figure 102) have served as the model for similar greenstone belts in Rhodesia, Australia, India, and Canada (*see* Viljoen and Viljoen 1969g; and Glikson 1976c). The volcanic and sedimentary successions of the Barberton Belt, shown in Figure 103, include the Onverwacht Group, 15 200 m thick composed mainly of komatiitic and tholeiitic lavas interlayered with some felsic calc-alkalic tuffs and tuff-breccias. Overlying the Onverwacht Group are 5300 m of sedimentary rocks belonging to the Fig Tree and Moodies Group. Some calc-alkalic tuffs and

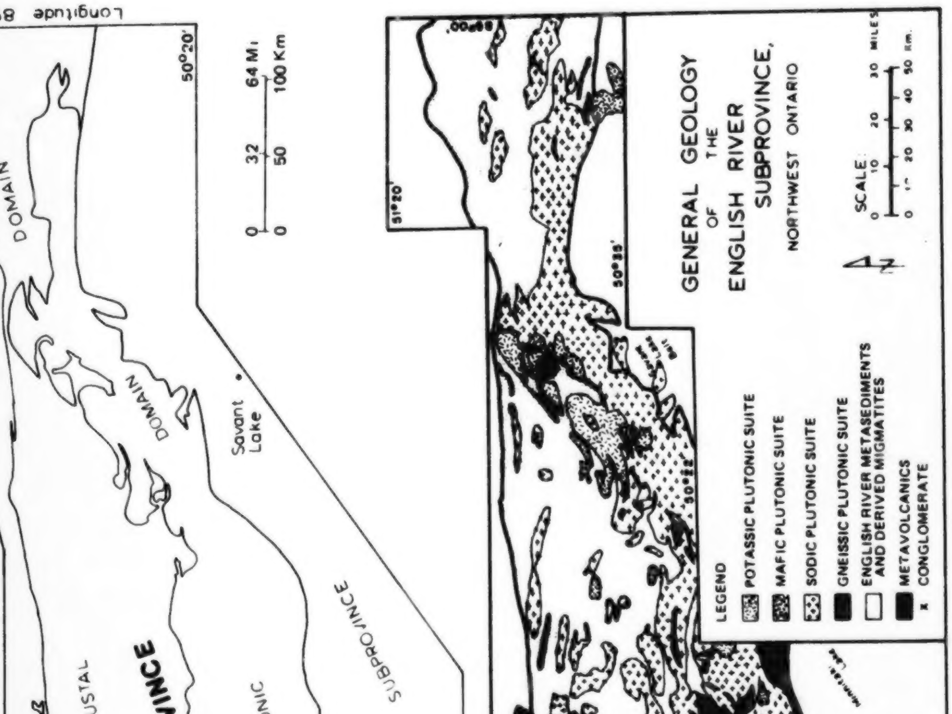


Figure 100. General geology of the English River Subprovince, northwestern Ontario (Breaks *et al.* 1978).

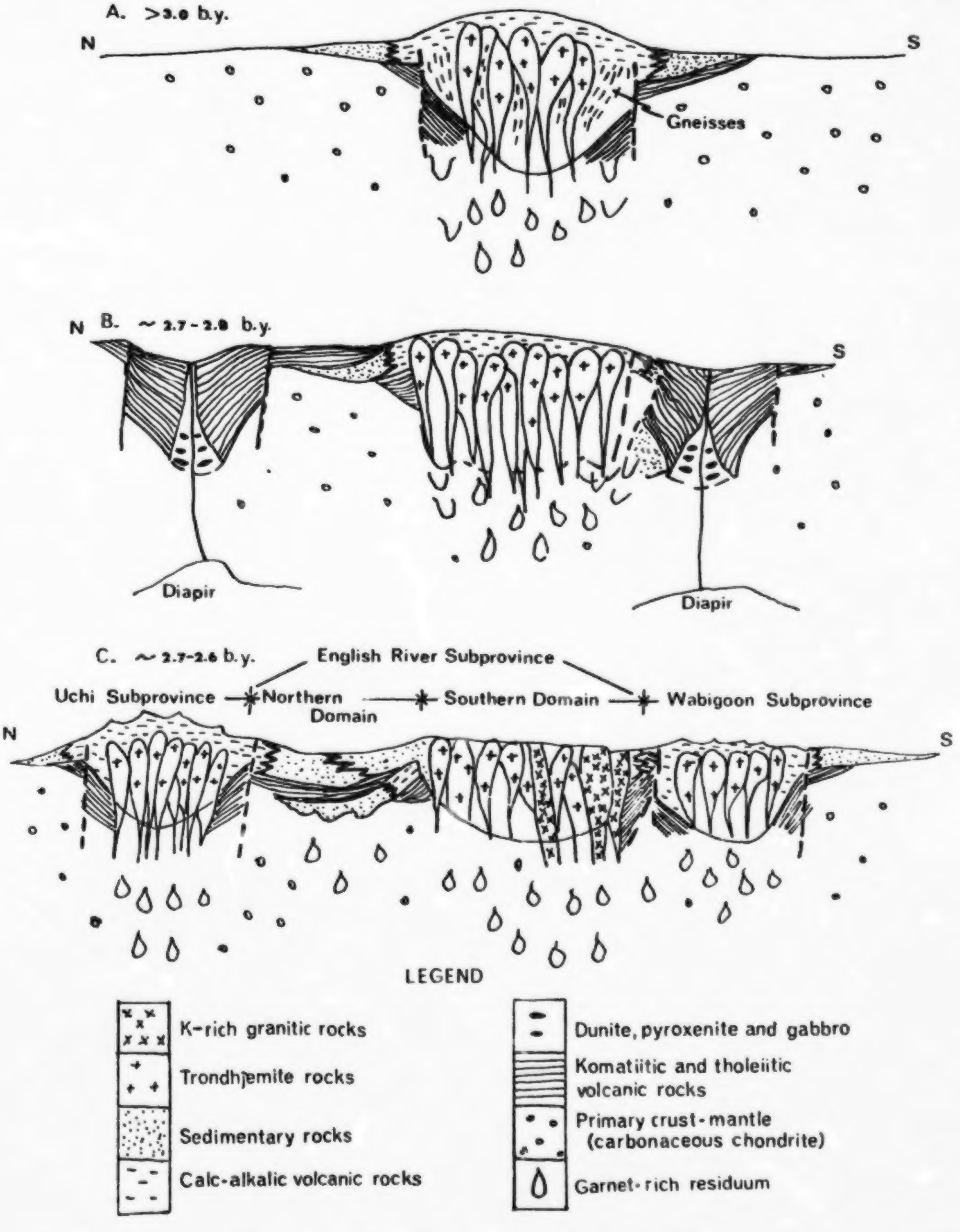


Figure 101. Diagram showing general development of the Uchi, English River, and Wabigoon Subprovinces of northwestern Ontario.

tuff-breccias are found near the top of the Fig Tree Group.

The Barberton Belt is about 100 km long and averages 30 km in width. Based on gravity studies, Darracott (1975) pointed out that in spite of the steeply dipping 15.2 km thick mafic volcanic succession, the Barberton Belt is less than 4 km deep, and is probably 'saucer'-shaped with a thick zone of sialic rock below it. In comparing the Barberton and Abitibi Belts, the Onverwacht Group corresponds closely to the Stoughton - Roquemaure and Larder Lake Groups which consist mainly of komatiitic and tholeiitic lavas interlayered with lesser amounts of felsic tuff and tuff-breccia. Unlike the Larder Lake, but like the Stoughton - Roquemaure Group, the Onverwacht Group is devoid of sedimentary rocks. Layered ul-

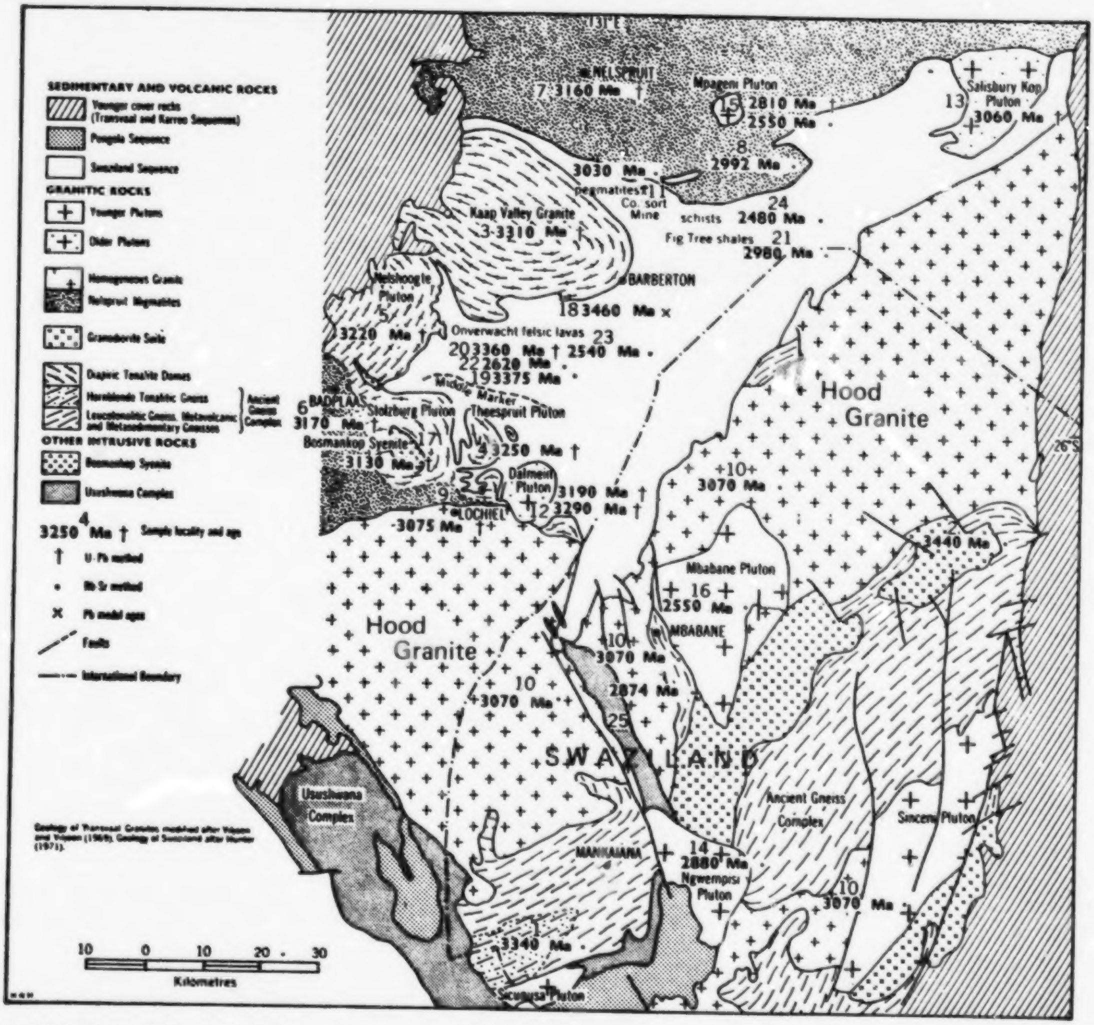


Figure 102. Simplified geological map of the eastern Transvaal and Swaziland showing the various granitic rock types surrounding the Barberton Mountain Land together with their respective age measurements. In addition, the localities of the most important age determinations carried out in the Barberton greenstone belt itself are illustrated (after Anhaeusser 1973).

tramafic complexes, composed of peridotite, pyroxenite, gabbro, norite, and anorthosite are concentrated on the northwestern side of the belt (Figure 104) which by analogy with the Stoughton - Roquemaure Group, suggests that komatiitic volcanism was more prevalent on the western and northwestern sides of the belt. Iron-rich gabbroic rocks do not appear to occur at the top of the layered complexes as they do in the Stoughton - Roquemaure Group. Only a few of the abundant basaltic lavas could classify as iron-rich tholeiitic basalts, have been noted in published rock analyses from the Onverwacht Group (for example, Viljoen and Viljoen 1969b, 1969c; Anhaeusser 1973).

Above the Onverwacht Group, there are no volcanic successions comparable to the Kinojevis and Blake River Groups. Instead, only a few calc-alkalic volcanic tuffs 550 m in thickness occur in the Schoongezicht Formation at the top of the Fig Tree Group which may have been a prism of sedimentary rocks and calc-alkalic tuffs (analogous to the Porcupine and Hunter Mine Groups). These tuffs were deposited on the edge of a calc-alkalic volcanic pile.

The Moodies Group of the Barberton Greenstone Belt can best be compared to the Timiskaming Group and Destor - Porcupine Complex in the Abitibi Belt (Eriksson 1978). However, granitic rocks appear to have provided detritus for the Moodies Group, while alkalic volcanism was responsible for much of the clastic material of the Timiskaming Group.

The Swaziland tonalitic migmatites and gneisses to the southeast of the Barberton Greenstone Belt are considered by Hunter (1973, 1974) to be older than the lowest units within the belt, whereas the opposite view is held by Viljoen and Viljoen (1969a, 1969b,

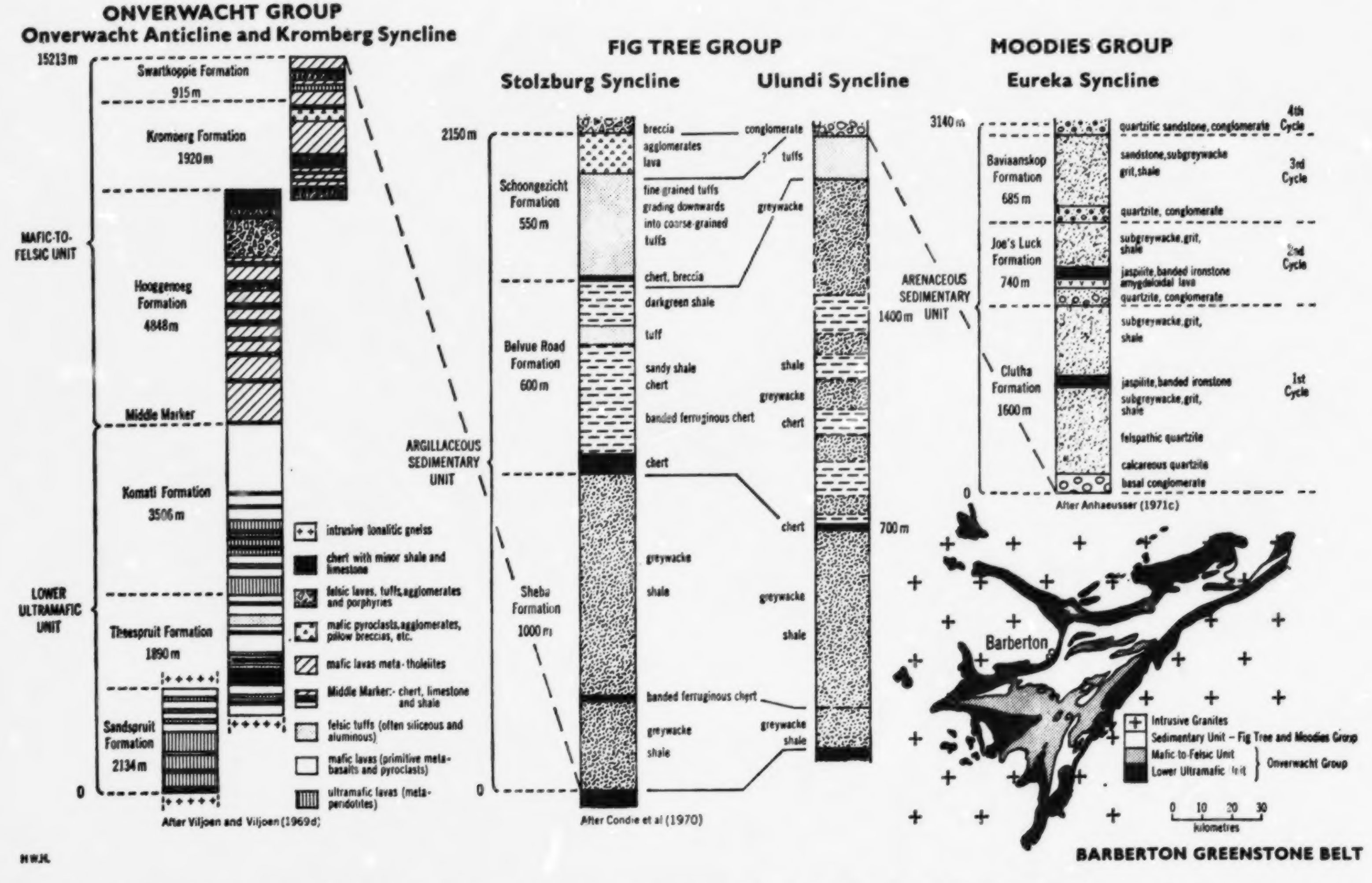

ONVERWACHT GROUP
Onverwacht Anticline and Kromberg Syncline
15213m
Swartkoppie Formation 915m
Kromberg Formation 1920m
MAFIC-TO-FELSIC UNIT
Hooggenoeg Formation 4848m
Middle Marker
Komati Formation 3506m
LOWER ULTRAMAFIC UNIT
Theespruit Formation 1890m
Sandspruit Formation 2134m
0
After Viljoen and Viljoen (1969d)
intrusive tonalitic gneiss
chert with minor shale and limestone
felsic lavas, tuffs, agglomerates and porphyries
mafic pyroclasts, agglomerates, pillow breccias, etc.
mafic lavas meta-tholeiites
Middle Marker:- chert, limestone and shale
felsic tuffs (often siliceous and aluminous)
mafic lavas (primitive meta-basalts and pyroclasts)
ultramafic lavas (meta-peridotites)
ARGILLACEOUS SEDIMENTARY UNIT
FIG TREE GROUP
Stolzburg Syncline
Ulundi Syncline
2150m
Schoongezicht Formation 550m
Belvue Road Formation 600m
Sheba Formation 1000m
0
breccia
agglomerates
lava
fine grained tuffs grading downwards into coarse-grained tuffs
chert, breccia
darkgreen shale
tuff
sandy shale
chert
banded ferruginous chert
chert
greywacke
shale
banded ferruginous chert
greywacke
shale
conglomerate
?
tuffs
greywacke
shale
greywacke
chert
chert
greywacke
shale
greywacke
shale
After Condie et al (1970)
ARENACEOUS SEDIMENTARY UNIT
1400m
700m
MOODIES GROUP
Eureka Syncline
3140m
Baviaanskop Formation 685m
Joe's Luck Formation 740m
Clutha Formation 1600m
0
quartzitic sandstone, conglomerate
sandstone, subgreywacke grit, shale
quartzite, conglomerate
subgreywacke, grit, shale
jaspilite, banded ironstone amygdaloidal lava
quartzite, conglomerate
subgreywacke, grit, shale
jaspilite, banded ironstone
subgreywacke, grit, shale
felspathic quartzite
calcareous quartzite
basal conglomerate
4th Cycle
3rd Cycle
2nd Cycle
1st Cycle
After Anhaeusser (1971c)
Barberton
Intrusive Granites
Sedimentary Unit – Fig Tree and Moodies Group
Mafic to Felsic Unit
Lower Ultramafic Unit
Onverwacht Group
0 10 20 30
kilometres
BARBERTON GREENSTONE BELT
H.W.H.

Figure 103. Stratigraphic sections depicting the various volcanic and sedimentary successions of the Swaziland Sequence in the Barberton Mountain Land, South Africa. The distribution of the lower ultramafic unit, the mafic-to-felsic unit, and the combined sedimentary units are shown in the inset general geological map of the Early Precambrian (≥3400 Ma old) Barberton greenstone belt (after Anhausser 1973).

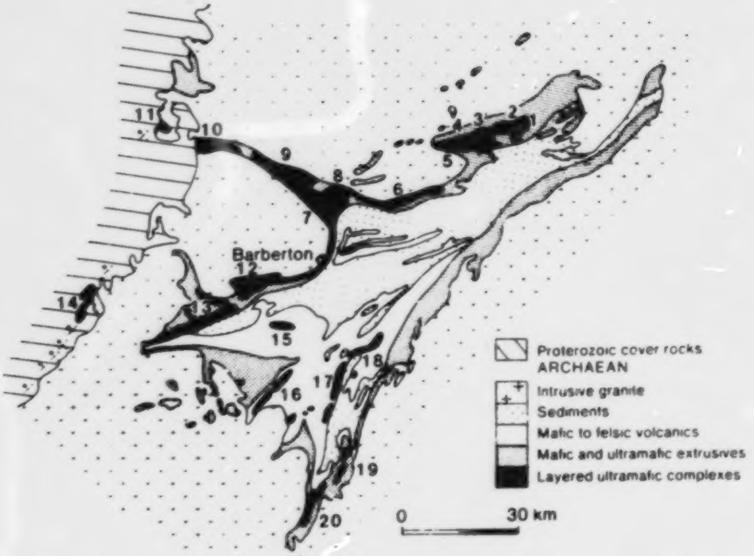


Figure 104. Geological map of the Barberton greenstone belt, showing the distribution of the layered differentiated ultramafic complexes in the Onverwacht volcanic sequence (after Anhaeusser 1975a, 1975b).

LAYERED ULTRAMAFIC COMPLEXES

1 KOEDOE
2 MAGNESITE
3 CENTRAL-CANAL
4 SHIP HILL
5 BUDD
6 SUGDEN
7 HANDSUP
8 MUNDT'S CONCESSION
9 HILLSIDE
10 KAAPSEHOOP
11 ELANDSHOEK
12 HILVERSUM
13 STOLZBURG
14 KALKKLOOF
15 GRANVILLE GROVE
16 ROSENTUIN
17 MSAULI
18 HAVELOCK
19 FORBES REEF
20 MOTJANE

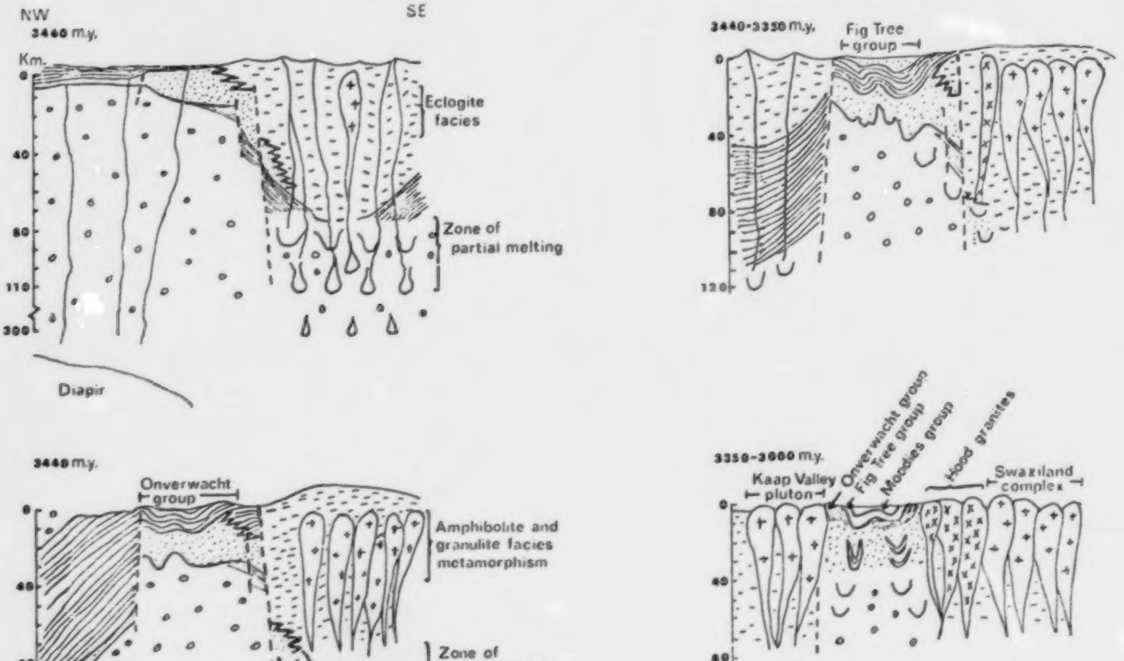


Figure 105. Diagram showing the development of the Barberton greenstone belt, South Africa by an adjacent megacauldron forming northeast of the primary megacauldron.

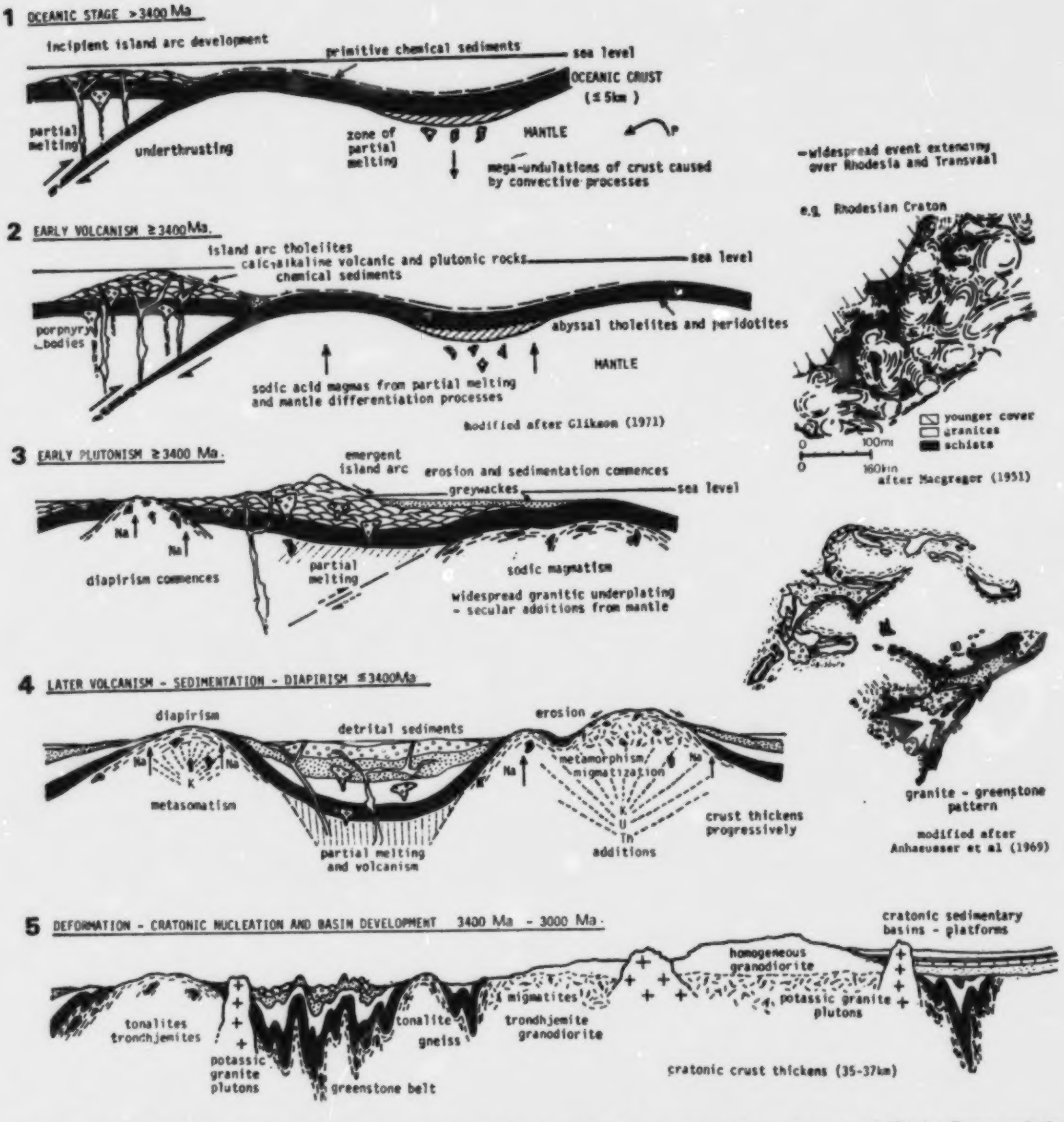

1 OCEANIC STAGE >3400 Ma
incipient island arc development
primitive chemical sediments
sea level
OCEANIC CRUST (≤ 5km)
partial melting
underthrusting
zone of partial melting
MANTLE
mega-undulations of crust caused by convective processes
2 EARLY VOLCANISM ≥3400 Ma.
island arc tholeiites
calc-alkaline volcanic and plutonic rocks
chemical sediments
sea level
porphyry bodies
abyssal tholeiites and peridotites
MANTLE
sodic acid magmas from partial melting and mantle differentiation processes
modified after Glikson (1971)
3 EARLY PLUTONISM ≥3400 Ma.
emergent island arc
erosion and sedimentation commences
greywackes
sea level
Na
diapirism commences
partial melting
sodic magmatism
widespread granitic underplating - secular additions from mantle
4 LATER VOLCANISM - SEDIMENTATION - DIAPIRISM ≤3400Ma
diapirism
detrital sediments
erosion
metasomatism
K
metamorphism, migmatization
K U Th additions
crust thickens progressively
partial melting and volcanism
5 DEFORMATION - CRATONIC NUCLEATION AND BASIN DEVELOPMENT 3400 Ma - 3000 Ma
cratonic sedimentary basins - platforms
homogeneous granodiorite
tonalites trondhjemites
potassic granite plutons
greenstone belt
tonalite gneiss
migmatites
trondhjemite granodiorite
potassic granite plutons
cratonic crust thickens (35-37km)
-widespread event extending over Rhodesia and Transvaal
e.g. Rhodesian Craton
younger cover
granites
schists
0 100mi
0 160km
after Macgregor (1951)
granite - greenstone pattern
modified after Anhaeusser et al (1969)

Figure 106. Schematic cross-sectional sketches illustrating the progressive stages of Early Precambrian crustal development in southern Africa. Process involving partial melting and granite underplating result in successive stages of transformation of a primary oceanic crust into regions having island arc-like characteristics, the latter finally nucleating to form stable cratons or shield areas (after Anhaeusser 1973).

1969c, 1969d, 1969e, 1969f, 1969g) and Anhaeusser (1973, 1975a, 1975b). Younger K-rich granitic rocks referred to as the Hood Granites separate the 2 groups of rocks (*see* Figure 102). Age determinations suggest that the migmatites are about the same age as the oldest volcanic rocks in the Barberton Belt.

The application of the model suggests that the Swaziland granitic rocks represent the earliest simato-sial transformation in this area. Therefore, a large calc-alkalic volcanic pile was likely present, with thick prisms of sedimentary rocks extending northwest at the time when this area was being flooded by komatiitic lavas. The lavas overlapped the sedimentary rocks as shown in Figure 105 which illustrates the gradual formation of the Barberton Greenstone Belt. The sial rocks below the volcanic ones may be sedimentary rock and felsic calc-alkalic tuffs with amphibolite and granulite facies metamorphism. The development of the Barberton Greenstone Belt shown in Figure 105 can be compared with that shown in Figure 106 by Anhaeusser (1973).

Summary

A study of the petrogenesis of the rocks in the Kirkland Lake part of the Abitibi Greenstone Belt reveals a panorama of Archean crustal development. Two cycles of komatiitic to tholeiitic to calc-alkalic volcanism are preserved, an older cycle represented by the Wabewawa, Catherine, and Skead Groups, and a younger cycle, by the Stoughton - Roquemaure (Larder Lake), Kinojevis, and Blake River Groups. The deposition of these rocks occurs among earlier formed calc-alkalic volcanic piles having cores replaced by Na-rich granitic rocks. As well, some Na-rich granitic rocks were emplaced in the core of the calc-alkalic pile formed by the Blake River Group. Therefore, each volcanic cycle can be thought of as the primary steps in the formation of sial crust and, where the calc-alkalic and sedimentary rocks are largely replaced by granitic rocks, the sima-to-sial transformation has become complete.

This transformation can be thought of as a megacauldron having 3 stages:

1. the formation of a thick mafic to ultramafic crust composed of lavas and intrusions on a primary crust of carbonaceous chondrite composition above an ascending diapir
2. the building of large calc-alkalic volcanic piles surrounded by sedimentary rocks from magmas generated by the partial melting of deeply buried mafic to ultramafic rocks converted to eclogite
3. the development of a granitic crust caused by the partial melting of deeply buried calc-alkalic and sedimentary rocks transformed into garnet bearing amphibolite or eclogite

A study of the distribution and relative ages of the rock-types in the Kirkland Lake area indicates that this 3-stage sima-to-sial transformation came about through the downward faulting, buckling, and tilting of crustal rocks occurring simultaneously with their replenishment by migrating upward and cooling at surface. To explain this, one can envisage megacauldrons in which subsidence of the crustal rocks is initiated by a thick mafic crust forming above a large mantle diapir. As the crustal rocks sink they are converted to eclogite, thus perpetuating the subsidence until most of the rocks at the surface are granitic rocks of low specific gravity. In the Abitibi Belt, although formation of granitic rocks had begun in the Blake River Group, this last stage was not extensive. Hence, much of the volcanic rock is still preserved so that the different stages of the sima-to-sial transformation in the youngest megacauldron can still be observed and this rock succession serves as a standard along with the model when studying other greenstone belts with less well preserved volcanic successions.

The model and the volcanic succession of the Abitibi Belt in the Kirkland Lake area when applied to other Archean greenstone belts indicates that most of them have resulted by the preservation of ultramafic and mafic volcanic rocks between areas in which the sima-to-sial transformation was nearing or at completion. This preservation took place because the volcanic rocks of one megacauldron were deposited on sedimentary and felsic calc-alkalic rocks formed peripherally to a more advanced megacauldron. Buoyed by these lighter rocks, the mafic and ultramafic rocks escaped burial and sinking, and became surrounded by granitic rocks.

Applying the model to other areas seems to indicate that the Archean crust formed in increments until continuous plates of crust, mainly of sialic composition resulted, and that the younger crustal additions tended to be smaller and thinner. The model supports concepts that the Earth was formed by cold, accreted material and that its mantle has since become warmer.

In future studies of Archean granite-greenstone terrains, the model could have a predictive value in determining the presence and location of rock types and tectonic features which may have mineral potential. However, trace element studies and age determinations are urgently needed, particularly of the intrusive rocks, to confirm the foregoing model.

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MARGINAL NOTES

LOCATION AND ACCESS

The Lightning River area is located about 50 km east of Matheson and 50 km north-northeast of Kirkland Lake, Ontario, on the boundary between Ontario and Quebec. The area encompasses Rand, Lamplugh, Stoughton, and Frecheville Townships as well as parts of Garrison, Holloway, Harker, and Marriott Townships, Indian Reserve No. 70, and Upper Lake Abitibi. The main access is via Highway 101 which transects the south part of the area. Logging roads extend off this highway and provide much secondary access to the south half of the area. The north half is accessible by boat which can be launched at the bridge where Highway 101 crosses the Ghost River. Air access is hampered by the generally turbid and shallow waters of Lake Abitibi.

MINERAL EXPLORATION

Mineral exploration started with the discovery of gold in 1906 along the south shore of Lake Abitibi (Miller 1907), and in 1917 in Harker and Holloway Townships (Knight *et al.* 1919). Documentation of this early mineral exploration can be found in a report by Gledhill (1925), and in the reports on Garrison, Harker, and Holloway Townships by Satterly (1949, 1951, 1953 respectively). A summary of the mineral exploration in Marriott and Stoughton Townships is given on Preliminary Maps P.823 and P.824 (Jensen 1973 b,c). In recent years mineral exploration has shifted to include asbestos and base metals as well as gold.

The search for asbestos has been concentrated in the ultramafic rocks that occur in the north parts of Garrison, Harker, and Holloway Townships and in the south parts of Lamplugh and Frecheville Townships. In 1950, Dominion Gulf Company and the Asbestos Corporation drilled 14 holes totalling 8 806 feet in the north part of Holloway Township and the southwest part of Frecheville Township following a magnetometer survey in the area (Resident Geologist's files, Ontario Ministry of Natural Resources, Kirkland Lake). In 1960, Canadian Johns-Manville Company Limited diamond drilled 7 holes totalling 3 391 feet in the southwest part of Frecheville and southeast part of Lamplugh Township (Resident Geologist's files, Ontario Ministry of Natural Resources, Kirkland Lake). Locations of this and other exploration for asbestos is shown on Ontario Division of Mines, Preliminary Maps P.797, P.798, and P.799 (Lovell and Frey 1973a,b,c).

Between 1968 and 1970, Canadian Johns-Manville Company Limited explored asbestos mineralization in the north part of Garrison Township. Underground exploration of 13 560 feet of drifting and cross-cutting and extensive diamond drilling was done with the stockpiling of some 20 000 tons of ore on surface for fibre quality tests (ODMNA 1970).

Only a limited amount of work has been done for base metals. In 1946, Hans Lundberg had an electromagnetic survey carried out in the north-central part of Rand Township and in 1964, Keevil Mining Group Limited conducted an airborne electromagnetic survey over the southwest parts of Rand and Lamplugh Townships. No conductors were found and the properties were consequently dropped (Resident Geologist's files, Ontario Ministry of Natural Resources, Kirkland Lake).

In 1973, Sciminex Limited and Anatole Resources Limited carried out a diamond drilling program consisting of four holes totalling about 2 000 feet in search of gold mineralization on 18 claims in the north part of Harker Township (Northern Miner 1973; L. Cunningham, Consultant, Kirkland Lake, personal communication 1973).

GENERAL GEOLOGY AND STRATIGRAPHY

Except for Keweenawan diabase dikes, all the bedrock is of Early Precambrian (Archean) age. A map of the stratigraphy and a table of stratigraphic units are shown in Figure 2. Sample locations of metavolcanics for which whole rock chemical analyses are available, are shown in Figure 3.

The oldest rocks are calc-alkalic basalts, andesites, dacites, and rhyolites called the Hunter Mine Group. These rocks occur at the west end of Upper Lake Abitibi, south parts of Indian Reserve No. 70, Rand, Lam-

The Stoughton-Roquemaure Group is conformably overlain by iron-enriched tholeiitic lavas referred to as the Kinojevis Group. At the top of the 10 km thick type-section of the Stoughton-Roquemaure Group, numerous layers of finely bedded calc-alkalic felsic tuff-breccias, tuffs, cherts, argillites, graphitic sediments, and ironstone appear in the metavolcanic succession with the tholeiitic lavas. Komatiitic lavas disappear from the succession and the lavas show a pronounced iron-enrichment in the upper 5 km thick metavolcanic succession of the 15 km thick southwest facing monoclinal succession forming the northeast side of the triangular syncline in Frecheville Township. Upward in the Kinojevis Group, the bedded tuffs and sediments decrease toward the center of the triangular syncline.

The tholeiitic lavas of the Kinojevis Group can be traced into the south part of Stoughton Township where they cross the Destor-Porcupine Fault Zone and can be followed westward south of the Destor-Porcupine Fault Zone. South of the fault zone, the Kinojevis Group attains a thickness greater than 10 km and is overlain by calc-alkalic metavolcanics belonging to the Blake River Group.

The Lake Abitibi Batholith is a composite trondhjemitic intrusion of massive to gneissic texture. Rocks throughout the batholith have chemical compositions equivalent to calc-alkalic dacite and rhyolite which suggests large volumes of komatiitic and tholeiitic lavas were not assimilated during the emplacement of the batholith. It is likely only calc-alkalic dacites and rhyolites were assimilated in significant amounts and it is probable that the partial melting of calc-alkalic metavolcanics at depth were responsible for the trondhjemitic magmas.

Along the south part of the main Destor-Porcupine Fault Zone, stocks and dikes of syenite, syenodiorite, and quartz-monzonite intrude the Kinojevis Group and the fault-bounded wedges of metasediments, alkalic and komatiitic metavolcanics. These intrusive rocks are absent north of the Destor-Porcupine Fault Zone as are the alkalic metavolcanics.

The pervasive grade of metamorphism in the metavolcanics is lower greenschist facies to sub-greenschist facies (prehnite and pumpellyite). Toward the margins of the Abitibi Batholith, middle and upper greenschist facies rank are attained, and amphibolite facies rocks are found at the contact. The low grade of metamorphism in the metavolcanics suggest that they did not undergo deep burial, and because of the short distance between the low-grade isograd and the edge of the batholith, the trondhjemitic magmas were emplaced at shallow depths.

Granulite facies metamorphic rank occurs in the metavolcanics at their contacts with the larger syenitic intrusions. Here, as well, the metamorphic grades decrease steeply away from the intrusive contacts. The generalized geology of this map area and the adjacent Magusi River map-area is shown In Figure 1.

STRUCTURAL GEOLOGY

The folding and the development of the Destor-Porcupine Fault Zone probably originated during the volcanism, sedimentation, and intrusion of the rocks of the area, rather than during a later period of tectonism. The grade of metamorphism does not indicate that the mafic lavas, which form the base of the metavolcanic succession in the Lightning River area, were ever buried to the depths equal to the thickness of this metavolcanic succession. As well, except at the margins of the batholith, the metavolcanics lack penetrative deformation generally associated with tectonism. It is more likely that the present attitudes of the lavas were largely developed during the deposition of succeeding lavas.

A topographically high calc-alkalic metavolcanic pile was probably present north of the main zone of the Destor-Porcupine Fault Zone and a sedimentary basin located to the south. Komatiitic magmatism developed in the basin covering the metasediments and lapped up onto the edges of this calc-alkalic pile.

North of the north branch of the Destor-Porcupine Fault Zone, as the komatiitic and tholeiitic lavas accumulated, they tilted southward as they subsided along the north branch fault zone. During tilting the lava flows became contained toward the fault, and particularly toward the center of the triangular syncline in Frecheville Township so that the lower lavas were not deeply buried. Calc-alkalic rocks below the komatiitic

in Lamplugh and Frecheville Townships. The only sizeable deposit of asbestos found to date is that in the north part of Garrison Township where Canadian Johns-Manville Company Limited stockpiled about 20 000 tons of ore for fibre quality tests. Here, the fibre occurs predominantly in dunite both as cross-fibre up to one inch wide and slip-fibre up to several inches long.

The Destor-Porcupine Fault Zone appears to have been a major conduit for hydrous fluids from depth and may have caused the development of some of the asbestos mineralization. Away from the Destor-Porcupine Fault Zone, the ultramafic rocks are less altered. In addition to asbestos, magnesite, and talc are developed in the ultramafic rocks along the fault zone and warrant further exploration.

Copper

Minor amounts of chalcopyrite occur in the calc-alkalic 'rhyolitic complexes' of the Hunter Mine Group, at the west end of Upper Lake Abitibi, and in Rand Township. Here, it appears as finely disseminated grains with pyrite, pyrrhotite, and sparse sphalerite.

Chalcopyrite occurs with nodular pyrite in the graphitic sediments and tuffs of the Hunter Mine Group and the interflow tuffs and sediments at the base of the Kinojevis Group.

Much of the area underlain by calc-alkalic rock in Indian Reserve No. 70, Rand, Lamplugh, and Frecheville Townships is poorly exposed and remains relatively unexplored.

Gold

Gold mineralization occurs mainly in the townships of Garrison, Harker, and Holloway near the Destor-Porcupine Fault Zone and according to Satterly (1949; 1951, 1953), several types of gold mineralization occur in the area.

Gold occurs: 1) In sheared and fractured zones in metavolcanics, metasediments, and felsic intrusive rocks; 2) In mineralized felsic dikes which have been carbonatized, or silicified, with or without a stockwork of quartz veins; 3) In quartz veins, fillings, and stockworks.

Gold may have been deposited in the metasediments off the margins of the earlier formed calc-alkalic pile. Some of these rocks are still preserved as fault-bounded wedges along the Destor-Porcupine Fault Zone, but other metasediments were probably displaced very deeply where they may have undergone partial melting. The Destor-Porcupine fault Zone being a deep-seated structure, would thus provide a conduit to surface for gold-bearing hydrothermal solutions and magmas following from the partial melting.

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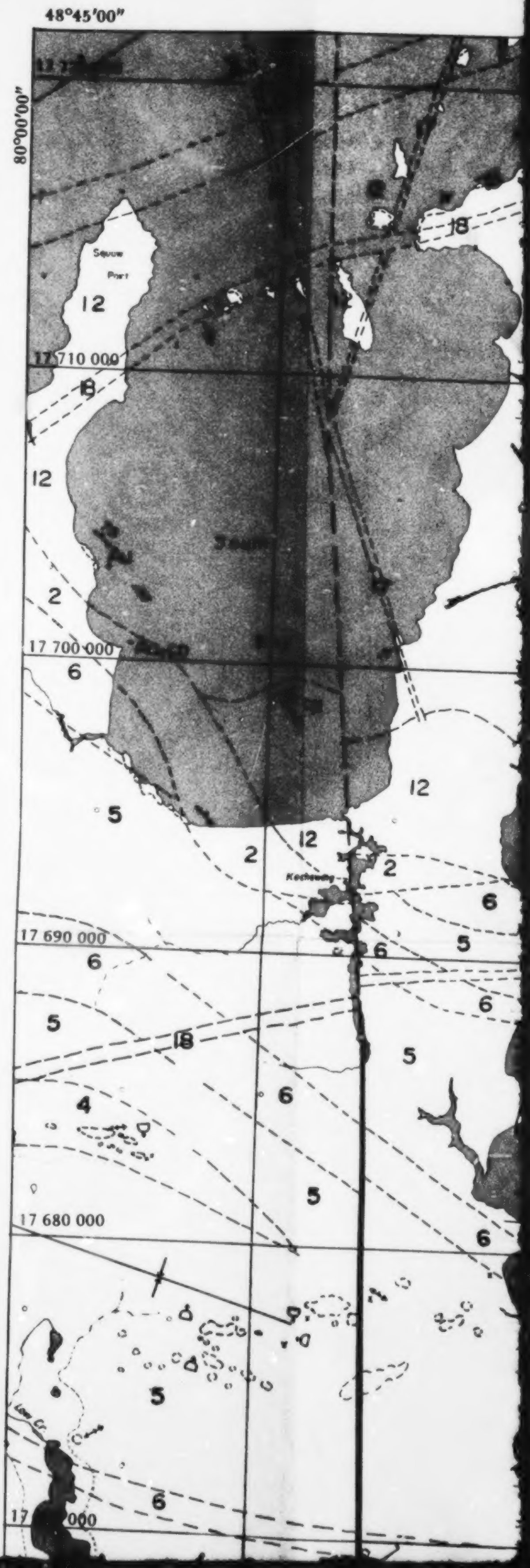

lle Townships. The only sizeable deposit of as-
t in the north part of Garrison Township where
Company Limited stockpiled about 20 000
tests. Here, the fibre occurs predominantly in
up to one inch wide and slip-fibre up to several

lt Zone appears to have been a major conduit
pth and may have caused the development of
neralization. Away from the Destor-Porcupine
rocks are less altered. In addition to asbestos,
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xploration.

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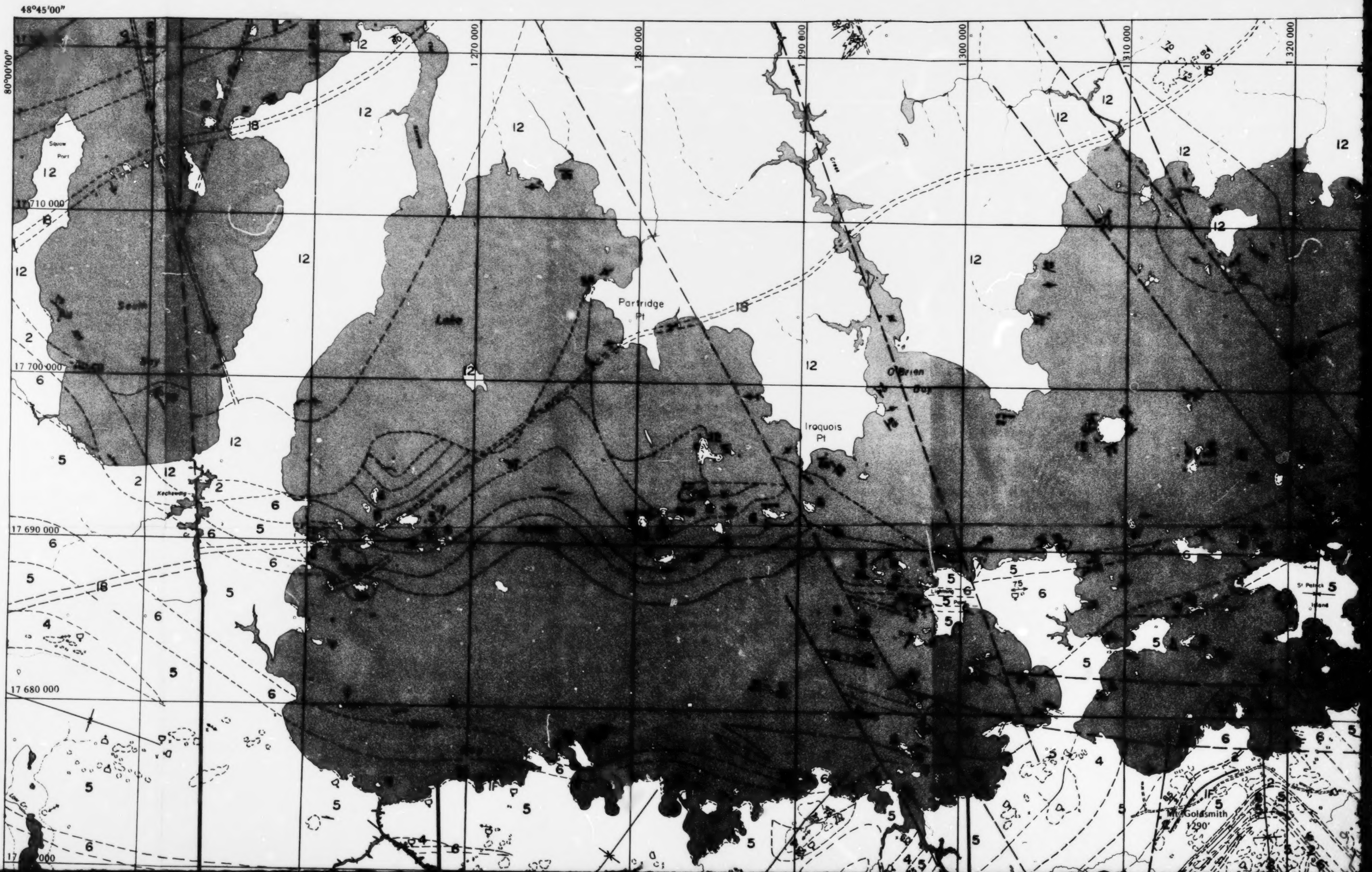

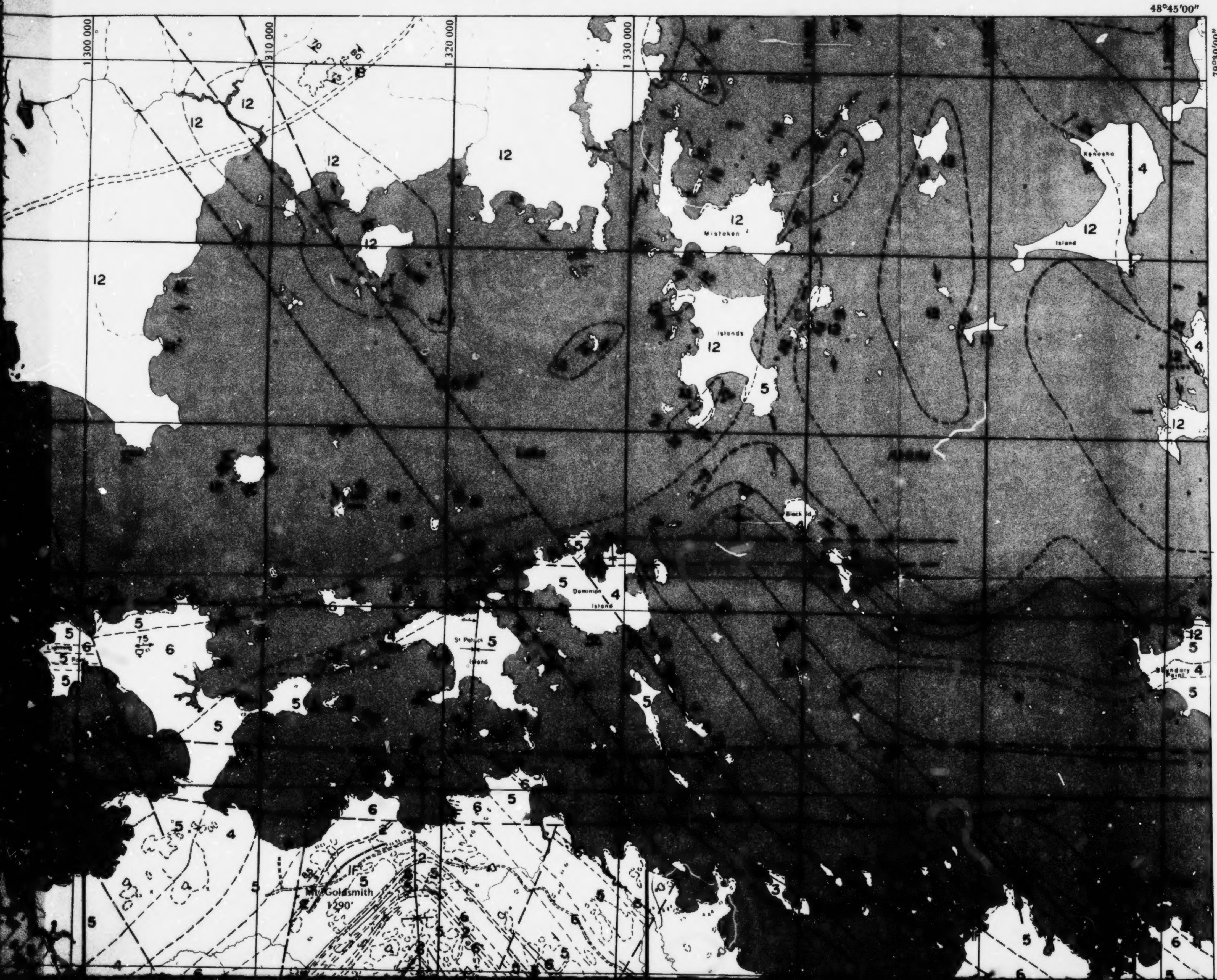

Ministry of Natural Resources
Hon. Alan W. Pope
Minister
W.T. Foster
Deputy Minister
Ontario

ONTARIO GEOLOGICAL SURVEY
MAP P.2433
GEOLOGICAL SERIES - PRELIMINARY MAP

PRECAMBRIAN GEOLOGY OF THE
LIGHTNING RIVER AREA
COCHRANE DISTRICT

Scale 1:63 360 or 1 inch to 1 mile

Miles 1 0 1 2 Miles

Kilometres 1 0 1 2 3 Kilometres

NTS Reference: 32 D/12
ODM-GSC Aeromagnetic Map: 45G
ODM Geological Compilation Map: 2205

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LEGEND

PHANEROZOIC
  CENOZOIC
    QUATERNARY
      PLEISTOCENE AND RECENT
        Till, reworked till, esker sand and gravel, varved clay, dune sand, alluvium, and peat

UNCONFORMITY

PRECAMBRIAN
  MIDDLE TO LATE PRECAMBRIAN (PROTEROZOIC)
    MAFIC INTRUSIVE ROCKS

18	18 Diabase, quartz diabase (Keweenawan)

ogist's files, Ontario Ministry of Natural Resources, Kirkland Lake). Locations of this and other exploration for asbestos is shown on Ontario Division of Mines, Preliminary Maps P.797, P.798, and P.799 (Lovell and Frey 1973a,b,c).

Between 1968 and 1970, Canadian Johns-Manville Company Limited explored asbestos mineralization in the north part of Garrison Township. Underground exploration of 13 560 feet of drifting and cross-cutting and extensive diamond drilling was done with the stockpiling of some 20 000 tons of ore on surface for fibre quality tests (ODMNA 1970).

Only a limited amount of work has been done for base metals. In 1946, Hans Lundberg had an electromagnetic survey carried out in the north-central part of Rand Township and in 1964, Keevil Mining Group Limited conducted an airborne electromagnetic survey over the southwest parts of Rand and Lamplugh Townships. No conductors were found and the properties were consequently dropped (Resident Geologist's files, Ontario Ministry of Natural Resources, Kirkland Lake).

In 1973, Sciminex Limited and Anatole Resources Limited carried out a diamond drilling program consisting of four holes totalling about 2 000 feet in search of gold mineralization on 18 claims in the north part of Harker Township (Northern Miner 1973; L. Cunningham, Consultant, Kirkland Lake, personal communication 1973).

GENERAL GEOLOGY AND STRATIGRAPHY

Except for Keweenawan diabase dikes, all the bedrock is of Early Precambrian (Archean) age. A map of the stratigraphy and a table of stratigraphic units are shown in Figure 2. Sample locations of metavolcanics for which whole rock chemical analyses are available, are shown in Figure 3.

The oldest rocks are calc-alkalic basalts, andesites, dacites, and rhyolites called the Hunter Mine Group. These rocks occur at the west end of Upper Lake Abitibi, south parts of Indian Reserve No. 70, Rand, Lamplugh, and Frecheville Townships and in Quebec, east of the map-area where they have been named. The Hunter Mine Group is characterized by 'rhyolite complexes' composed of breccias cut by numerous subvolcanic dikes of andesite, dacite, and rhyolite composition. These rocks contain feldspar and quartz phenocrysts. The complexes grade into bedded tuffs and tuff-breccias which in turn grade into cherts, iron formations, and in places, wacke. The facies changes in the Hunter Mine Group suggest a large calc-alkalic pile once existed in the vicinity of the Lake Abitibi Batholith. The rhyolite complex in Rand Township is surrounded by calc-alkalic basalt and andesite flows interlayered with tuff breccias of the same composition as well as dacite and rhyolite.

Overlying the Hunter Mine Group are komatiitic and tholeiitic lavas of the Stoughton-Roquemaure Group which is more than 10 km thick in its type-section. In the type-section it overlies the Hunter Mine Group in Roquemaure Township and forms a steeply southeast-dipping monoclinal succession, the upper part of which forms the bedrock in the northeast half of Stoughton Township. The upper part of this succession can be traced westward across Lake Abitibi where again, the lavas can be seen to overlie the Hunter Mine Group. Elsewhere, the Stoughton-Roquemaure Group is intruded by the Lake Abitibi Batholith toward its base.

The calc-alkalic metavolcanics of the Hunter Mine Group in the Lamplugh area are cut by stocks and sills of peridotite which may have been feeders for the komatiitic lavas. In the south part of Lamplugh Township, the calc-alkalic rocks are overlain by a thick, flat-lying fractionated komatiitic lava flow which may have been ponded on the irregular calc-alkalic metavolcanic topography. The flow consists of a massive basal peridotite layer overlain by pyroxenite and gabbro similar in composition to magnesium-rich tholeiitic basalt. At higher elevations, the flow is capped by a finely bedded, 30 cm thick unit of calc-alkalic dacite tuff overlain by thick massive flows of iron-rich tholeiitic basalt. A similar group of rocks occur in the north part of Garrison Township, except, here they are tipped steeply on their side to the north.

Fault-bounded wedges of komatiitic lava are also found along the Destor Porcupine Fault Zone and are considered as well, to be part of the Stoughton-Roquemaure Group.

Komatiitic volcanism forming the Stoughton-Roquemaure Group appears to have begun in the basin to the south and spread northward engulfing the calc-alkalic volcanic pile represented by the Hunter Mine Group.

gest that they did not undergo deep burial, and because of the short distance between the low-grade isograd and the edge of the batholith, the trondhjemitic magmas were emplaced at shallow depths.

Granulite facies metamorphic rank occurs in the metavolcanics at their contacts with the larger syenitic intrusions. Here, as well, the metamorphic grades decrease steeply away from the intrusive contacts. The generalized geology of this map area and the adjacent Magusi River map-area is shown in Figure 1.

STRUCTURAL GEOLOGY

The folding and the development of the Destor-Porcupine Fault Zone probably originated during the volcanism, sedimentation, and intrusion of the rocks of the area, rather than during a later period of tectonism. The grade of metamorphism does not indicate that the mafic lavas, which form the base of the metavolcanic succession in the Lightning River area, were ever buried to the depths equal to the thickness of this metavolcanic succession. As well, except at the margins of the batholith, the metavolcanics lack penetrative deformation generally associated with tectonism. It is more likely that the present attitudes of the lavas were largely developed during the deposition of succeeding lavas.

A topographically high calc-alkalic metavolcanic pile was probably present north of the main zone of the Destor-Porcupine Fault Zone and a sedimentary basin located to the south. Komatiitic magmatism developed in the basin covering the metasediments and lapped up onto the edges of this calc-alkalic pile.

North of the north branch of the Destor-Porcupine Fault Zone, as the komatiitic and tholeiitic lavas accumulated, they tilted southward as they subsided along the north branch fault zone. During tilting the lava flows became contained toward the fault, and particularly toward the center of the triangular syncline in Frecheville Township so that the lower lavas were not deeply buried. Calc-alkalic rocks below the komatiitic lavas subsided downward and northward as well. Rocks at the base of the calc-alkalic pile were probably converted to eclogite, sank and were partially melted to form trondhjemitic magmas which returned to surface forming the batholith and leaving the garnet-rich residues to sink deeper into the mantle. This would have caused the adjacent calc-alkalic rocks to be displaced downward and inward under the developing batholith to replace the 'processed' rock and making room for the mafic lavas to accumulate at surface. South of the north branch of the Destor Porcupine Fault Zone, the calc-alkalic metavolcanics were not displaced downward, but instead, remained stable and were not covered by an extensive accumulation of komatiitic lavas.

As the komatiitic and tholeiitic lavas of the Stoughton-Roquemaure Group and Kinojevis Group accumulated in the basin, south of the calc-alkalic pile, they subsided along a series of slip-planes now marked by the main zone of the Destor Porcupine Fault Zone; the downward displacement being 10 to 20 km according to the thickness of ultramafic and mafic lavas observed elsewhere. The only remaining portions of the Stoughton-Roquemaure Group are in fault bounded wedges of rock located in the fault zone. Many of these komatiites are strongly hydrated, carbonatized and alkali enriched and sheared as are the wedges of turbiditic metasediments with the komatiitic lavas.

Metasediments south of the calc-alkalic pile, were likely buried deeply enough south of the Destor-Porcupine Fault Zone that they underwent dehydration and partial melting with some of the lower komatiitic lavas. Hydrous solutions and syenitic magmas would be expected to penetrate the rocks at surface along the south part of the fault zone and to the south as observed.

In Stoughton and Marriott Townships, where the north branch of the Destor-Porcupine Fault Zone converges with its main zone, downward displacement of the rocks on both sides occurred so that the differential movement between the two sides of the fault is small and the expression of the fault is weak.

ECONOMIC GEOLOGY

Asbestos

Asbestos occurs in the serpentinized dunite and peridotite in Garrison, Harker, and Holloway Townships and has previously been described by Satterly (1951, 1953). Similar small amounts of fibre were also observed

Gold may have been deposited in the metasediments off the margins of the earlier formed calc-alkalic pile. Some of these rocks are still preserved as fault-bounded wedges along the Destor-Porcupine Fault Zone, but other metasediments were probably displaced very deeply where they may have undergone partial melting. The Destor-Porcupine fault Zone being a deep-seated structure, would thus provide a conduit to surface for gold-bearing hydrothermal solutions and magmas following from the partial melting.

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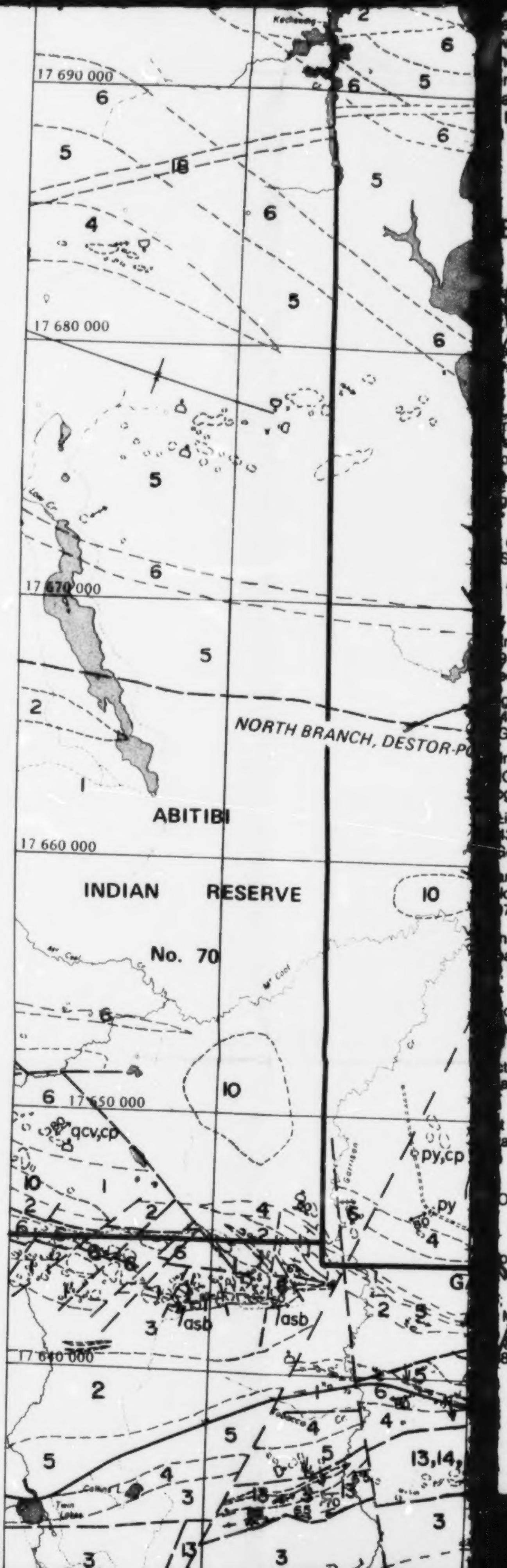

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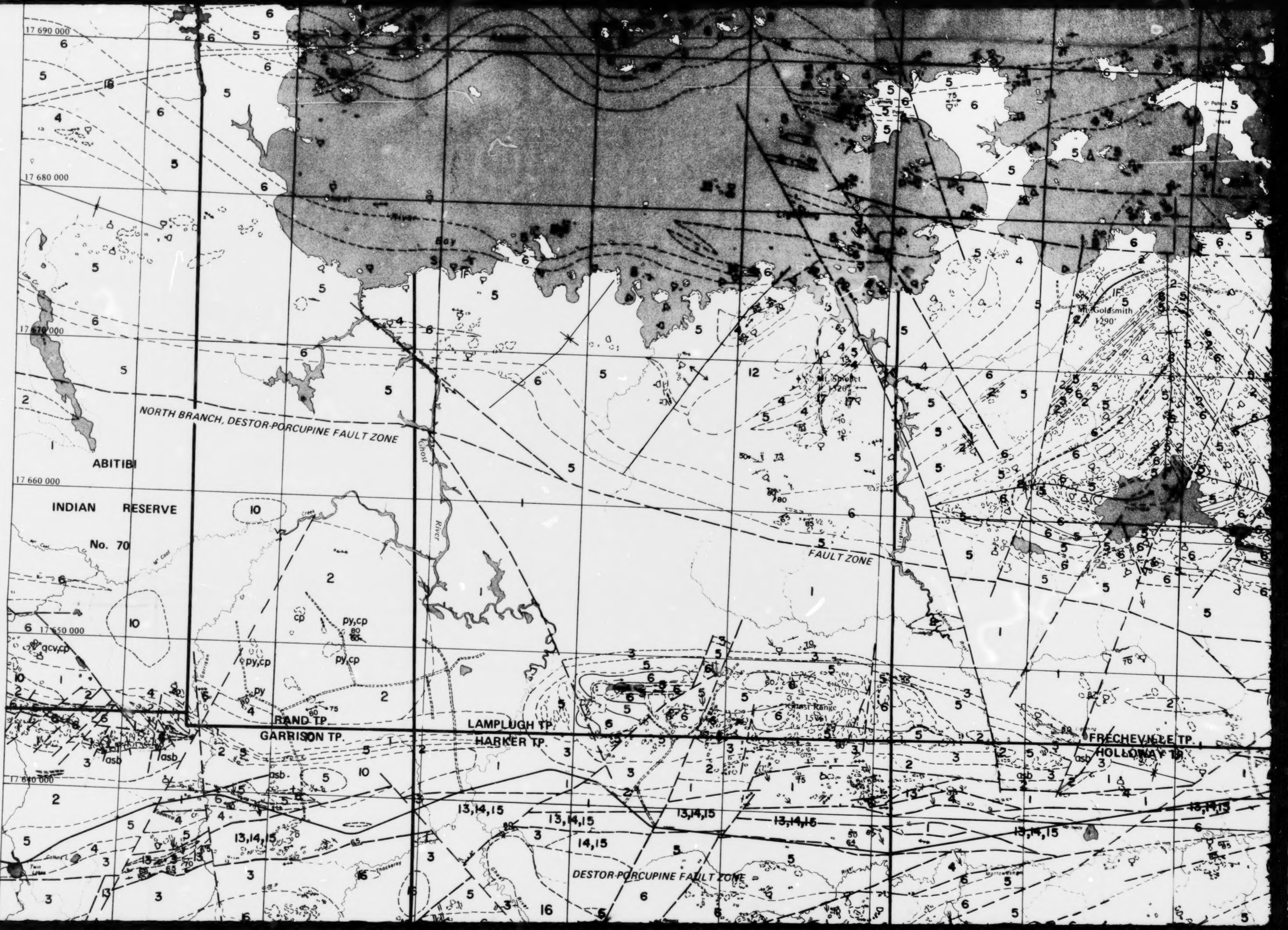

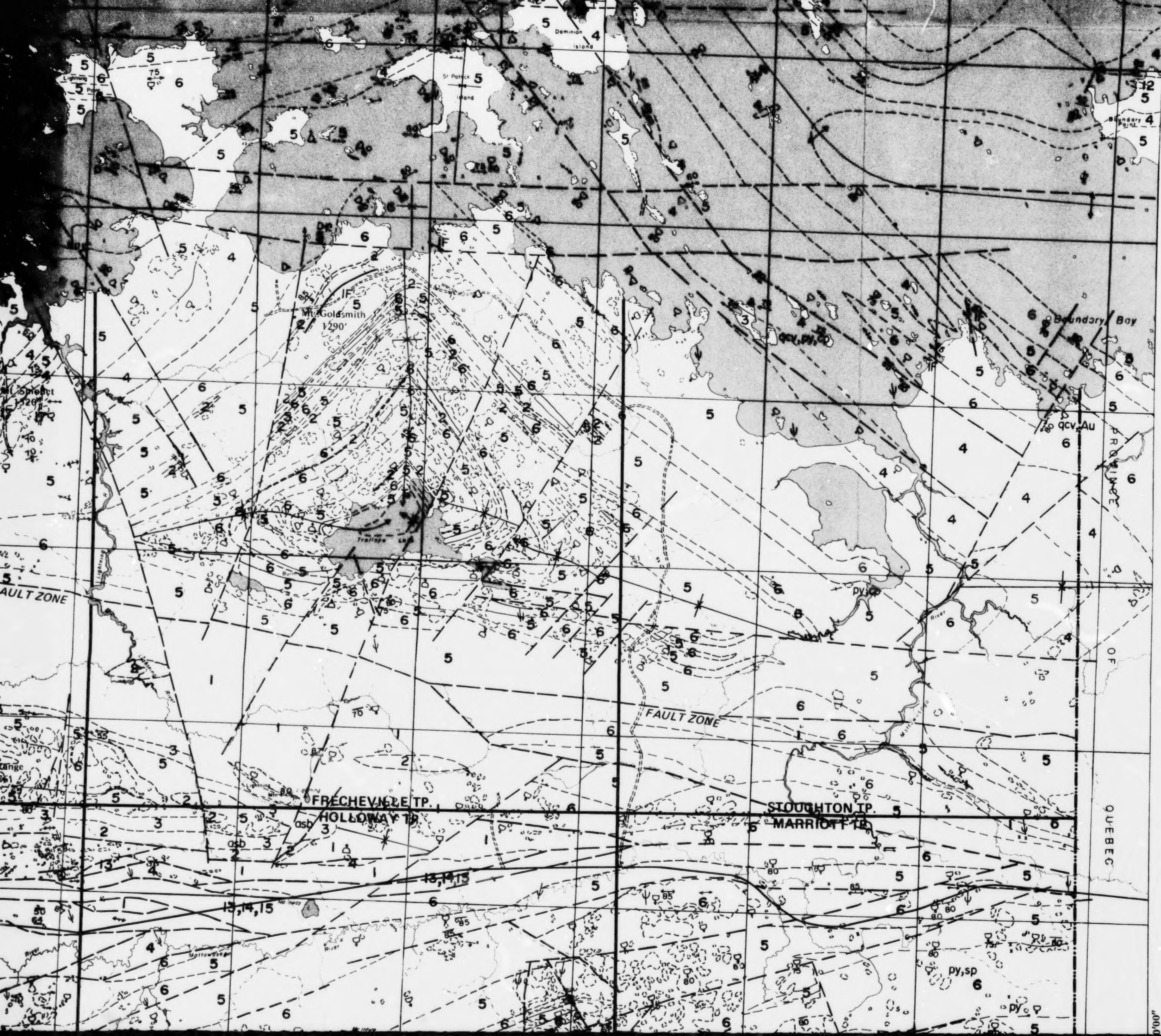

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LEGEND

PHANEROZOIC
CENOZOIC
QUATERNARY
PLEISTOCENE AND RECENT
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UNCONFORMITY

PRECAMBRIAN
MIDDLE TO LATE PRECAMBRIAN (PROTEROZOIC)
MAFIC INTRUSIVE ROCKS

18 18 Diabase, quartz diabase (Keweenawan)

INTRUSIVE CONTACT

EARLY PRECAMBRIAN (ARCHEAN)
MAFIC INTRUSIVE ROCKS

17 17 Diabase (Matachewan)

INTRUSIVE CONTACT

FELSIC INTRUSIVE ROCKS

16 16 Equigranular and porphyritic syenodiorite, monzonite, feldspar porphyry, pegmatite, lamprophyre

INTRUSIVE CONTACT

ALKALIC METAVOLCANICS
Felsic Metavolcanics

15 15 Sodic trachyte, benmorite, alkalic dacite, rhyolite

Mafic Metavolcanics

14 14 Hawaiite, alkali basalt, mugearite, nephelinite

METASEDIMENTS

13 13 Conglomerate, wacke, arkose, argillite, sandstone, ironstone

UNCONFORMITY

FELSIC INTRUSIVE ROCKS

12 12 Trondhjemite, quartz diorite, diorite

INTRUSIVE CONTACT

MAFIC INTRUSIVE ROCKS

11 11 Gabbro, quartz gabbro, hornblende gabbro, granophyre

INTRUSIVE CONTACT

ULTRAMAFIC INTRUSIVE ROCKS

10 10 Peridotite, dunite, pyroxenite, rodingite

INTRUSIVE CONTACT

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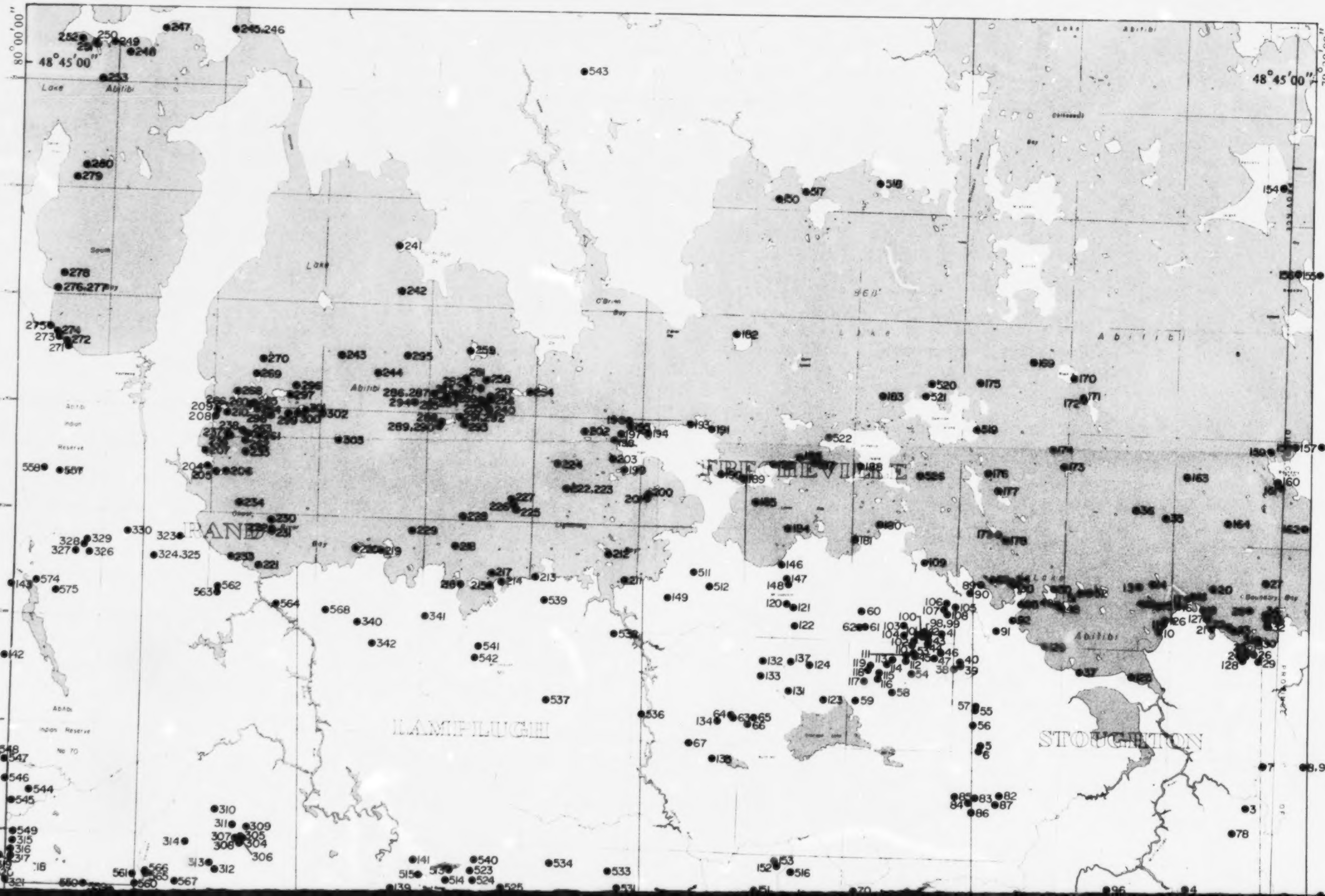

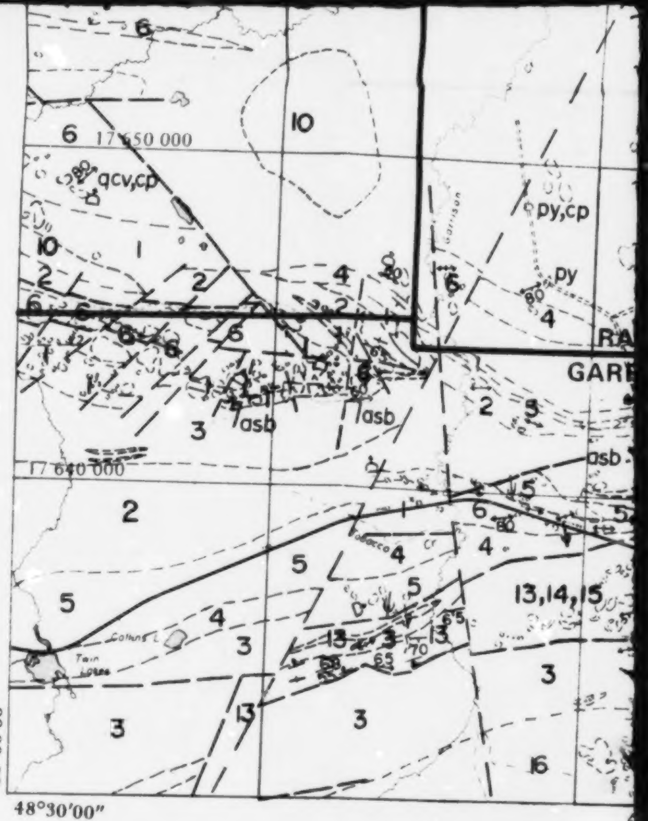

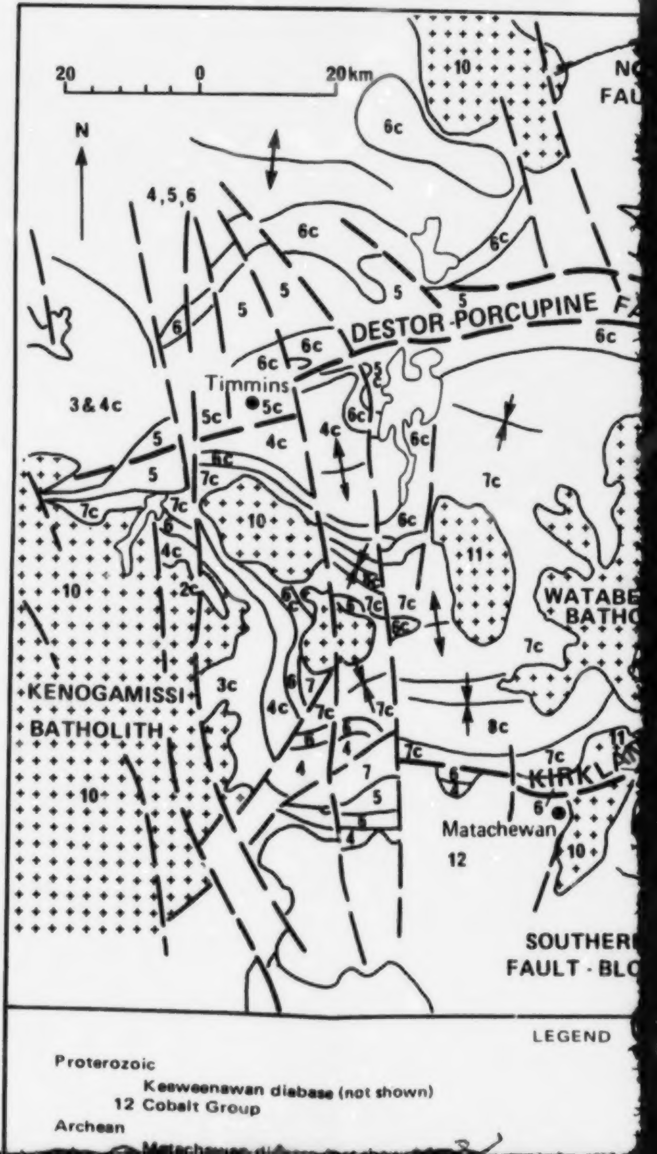

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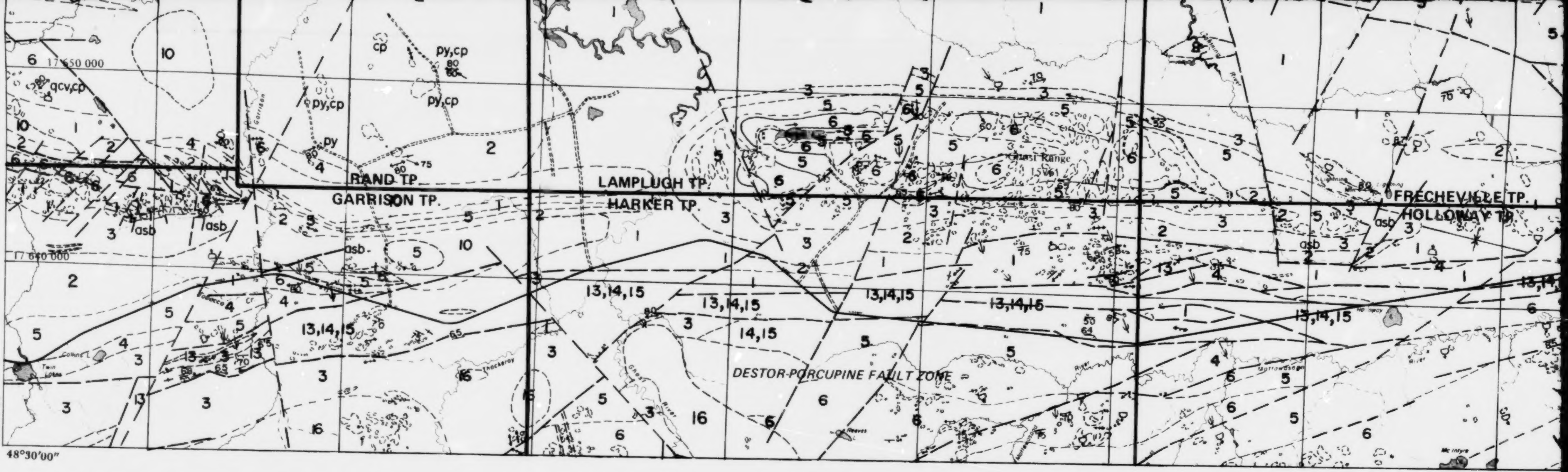
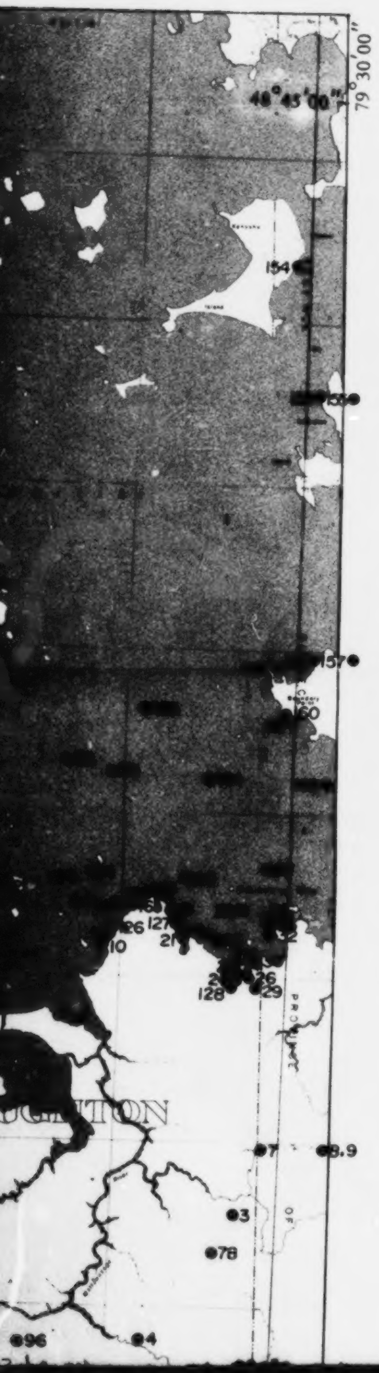
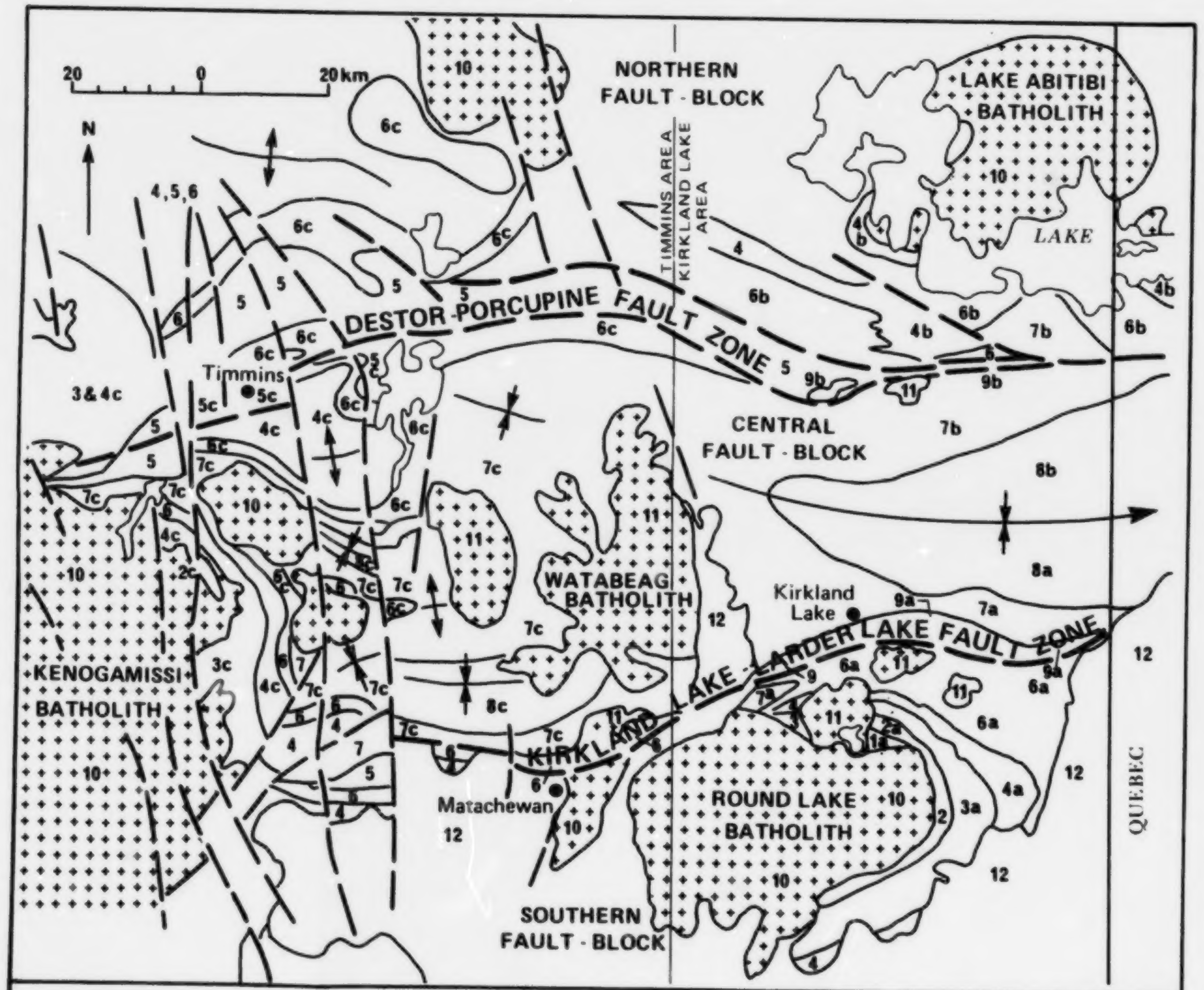
North Half of Holloway Township; Ontario Department of
, 1953, p.1-38. Accompanied by Map No.1953-4, Scale 1



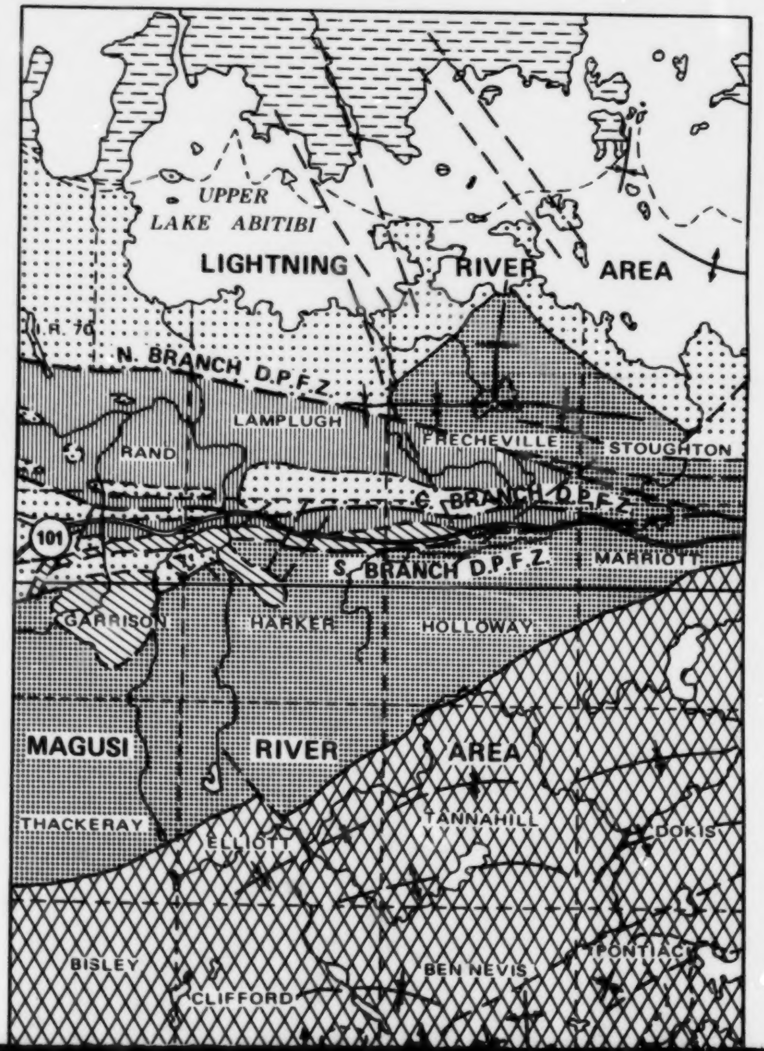




LEGEND

Proterozoic
Keeweenawan diabase (not shown)
12 Cobalt Group
Archean
Matachewan diabase (not shown)
Granitic rocks

7 7a, 7b, Kinojevis Group, 7c Kinojevis Group, (Middle Fm., Tisdale Group)
6 6a Larder Lake Group, 6b Stoughton Roquemaure Group, 6c Lower Fm., Tisdale Group



Fault

Syncline

Anticline

Conformable contact

Unconformable contact

Intrusive contact

Abitibi Batholith

Destor - Porcupine Complex

Blake River Group

Kinojevis Group

Stoughton - Roquemaure Group

LOCATI

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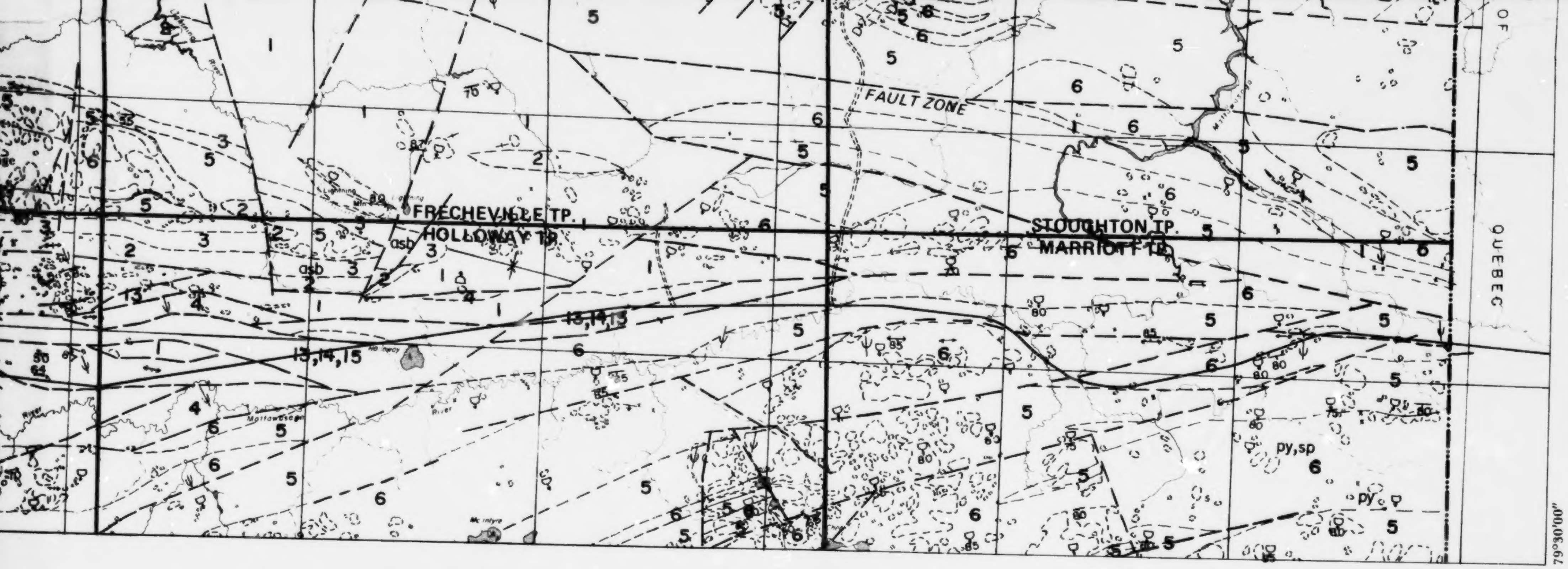
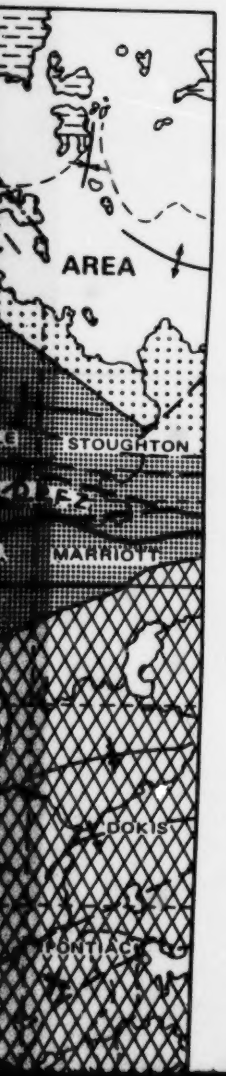
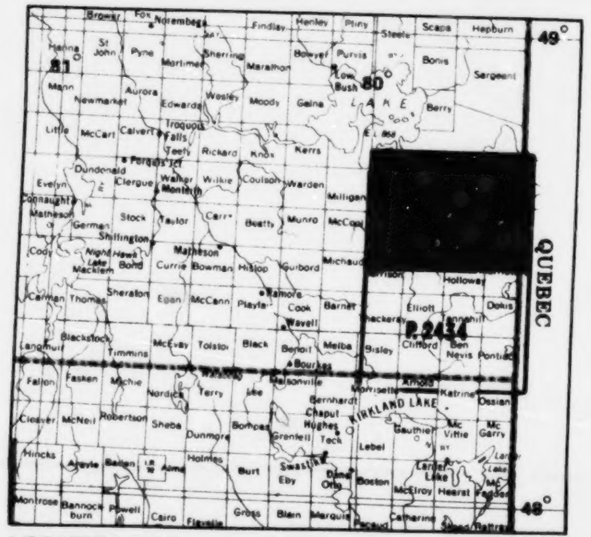




- Fault
- Syncline
- Anticline
- Conformable contact
- Unconformable contact
- Intrusive contact
- ▤ Abitibi Batholith
- ▧ Destor - Porcupine Complex
- ▦ Blake River Group
- ▩ Kinojevis Group
- ░ Stoughton - Roquemaure Group
- ▥ Hunter Mine Group



LOCATION MAP

Scale 1: 1 584 000
or 1 inch to 25 miles

SYMBOLS

- Glacial striae
- × Small bedrock outcrop
- Area of bedrock outcrop
- Bedding, top unknown; (inclined, vertical)
- Bedding, top indicated by arrow; (in-
- Schistosity; (horizontal, inclined, vertical)
- Gneissosity; (horizontal, inclined, vertical)
- Lineation with plunge
- Geological boundary, observed
- Geological boundary, position interpreted

METAL AND MINERAL ABBREVIATIONS

asb	Asbestos
Au	Gold
cp	Chalcopyrite
IF	Iron Formation
py	Pyrite
qcv	Quartz carbonate vein
sp	Sphalerite

SOURCES OF INFORMATION

Base map derived from Forest Resources Inventory maps, Lands and Waters Group, Ontario Ministry of Natural Resources.

Geology not tied to survey lines.

Resident Geologist's files, Ontario Ministry of Natural Resources, Kirkland Lake.

Magnetic declination 10.00°W in 1973.

CREDITS

Geology by L.S. Jensen and assistants, 1973.

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Issued 1982

Felsic Metavolcanics

15 — 15 Sodic trachyte, benmorite, alkalic dacite, rhyolite

Mafic Metavolcanics

14 — 14 Hawaiite, alkali basalt, mugearite, nephelinite

METASEDIMENTS

13 — 13 Conglomerate, wacke, arkose, argillite, sandstone, ironstone

UNCONFORMITY

FELSIC INTRUSIVE ROCKS

12 — 12 Trondhjemite, quartz diorite, diorite

INTRUSIVE CONTACT

MAFIC INTRUSIVE ROCKS

11 — 11 Gabbro, quartz gabbro, hornblende gabbro, granophyre

INTRUSIVE CONTACT

ULTRAMAFIC INTRUSIVE ROCKS

10 — 10 Peridotite, dunite, pyroxenite, rodingite

INTRUSIVE CONTACT

UPPER CALC-ALKALIC METAVOLCANICS

Felsic Metavolcanics

9 — 9 Massive flow breccia, tuff-breccia, crystal tuff, feldspar, and quartz feldspar porphyritic rhyolitic and dacitic rocks

Mafic and Intermediate Metavolcanics

8 — 8 Massive and pillowed flows, pillow-breccia, pyroclastic breccia, volcanoclastic, tuff, lapilli-tuff, amygdaloidal, porphyritic feldspar, basaltic and andesitic rocks

THOLEIITIC METAVOLCANICS

Felsic Metavolcanics

7 — 7 Spherulitic and granular tuff, tuff-breccia, cherty tuff, dacite, and rhyolite

Iron-Rich Mafic Metavolcanics

6 — 6 Black to dark green, massive, pillowed, pillow-breccia, hyaloclastic, variolitic, amygdaloidal basalt, andesite rocks, and interflow sediments

Magnesium-Rich Metavolcanics

5 — 5 Grey to green, massive, pillowed, pillow-breccia, hyaloclastic, variolitic, porphyritic feldspar, amygdaloidal basaltic rocks, and interflow sediments

KOMATIITIC METAVOLCANICS

Basaltic Metavolcanics

4 — 4 Massive, pillowed, pillow-breccia, hyaloclastic, variolitic, spinifex-textured basaltic rock

Ultramafic Metavolcanics

3 — 3 Massive, pillowed, polysutured, spinifex-textured ultramafic metavolcanics

UNCONFORMITY

LOWER CALC-ALKALIC METAVOLCANICS

Felsic Metavolcanics

2 — 2 Tuff-breccia, crystal tuff, tuff, quartz and feldspar porphyritic rhyolitic and dacitic rocks, argillite, chert, and ironstone

Mafic and Intermediate Meta

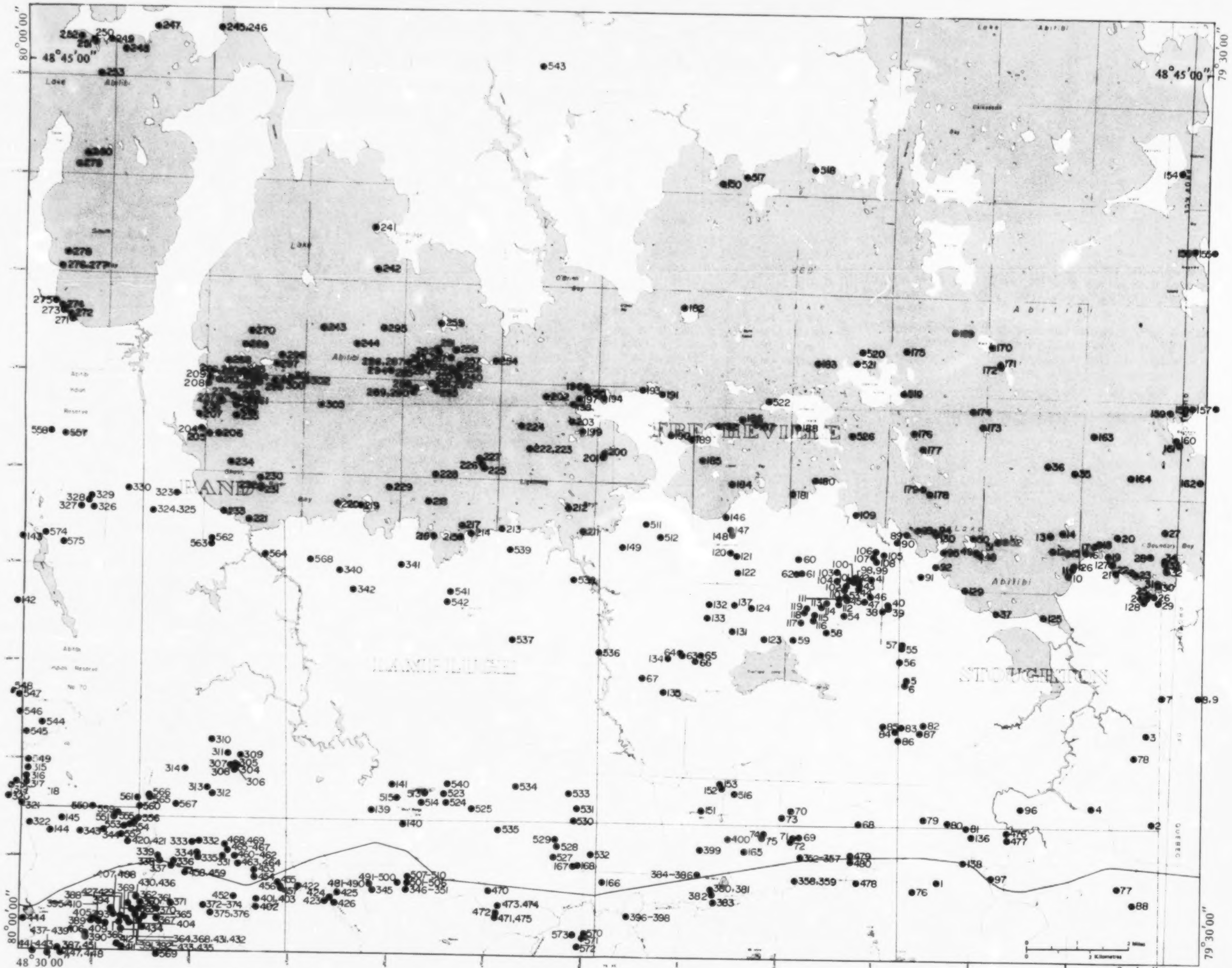
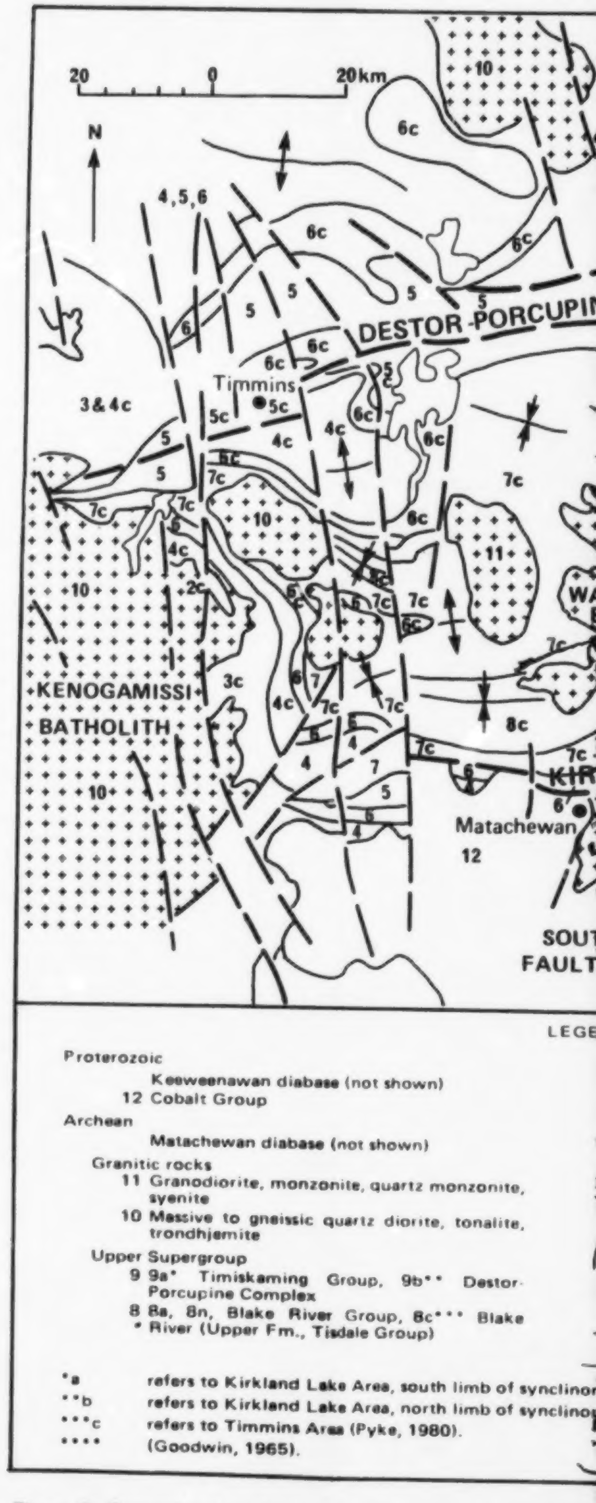


Figure 3: Sample locations for chemical analyses.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DESTOR-PORCUPINE
Timmins
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KENOGAMISSI BATHOLITH
Matachewan
12
SOUT
FAULT
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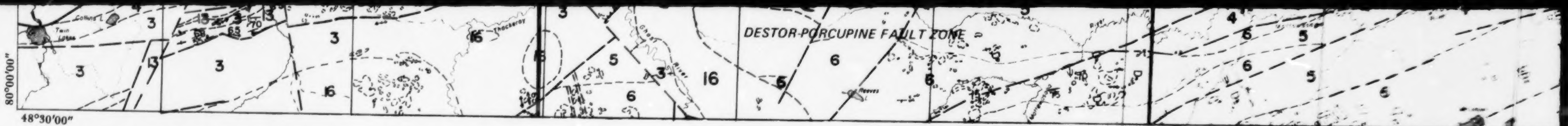
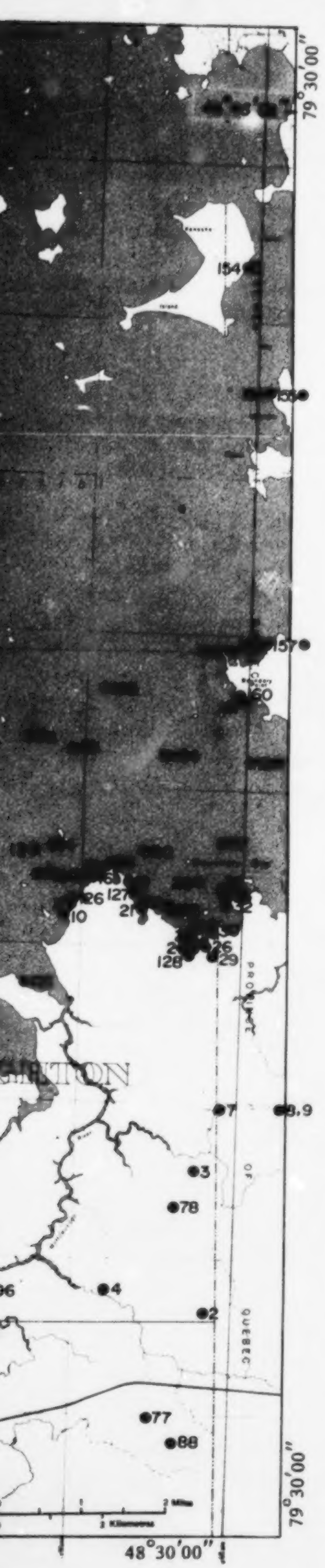
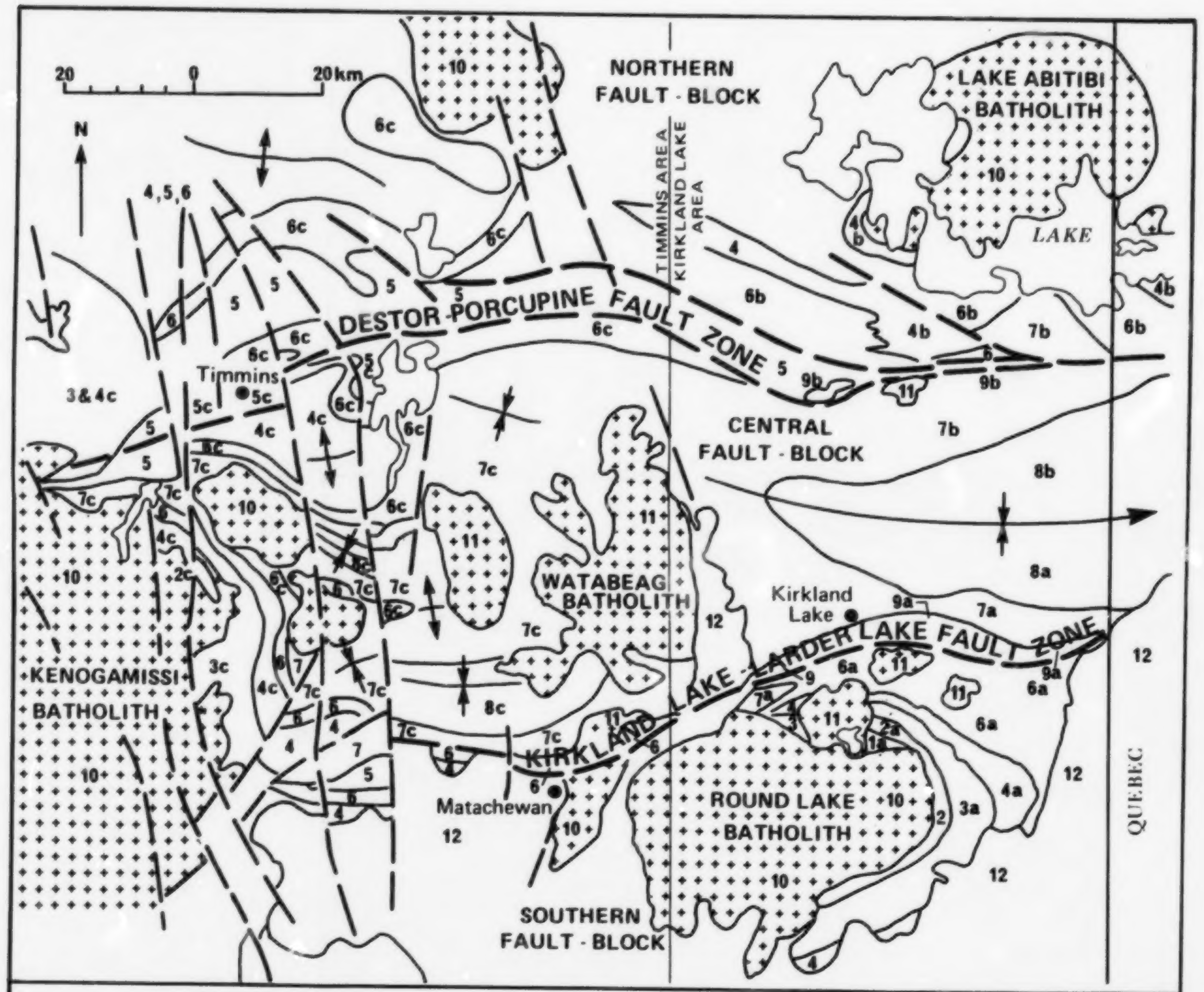
WA
LEGE
Proterozoic
Keeweenawan diabase (not shown)
12 Cobalt Group
Archean
Matachewan diabase (not shown)
Granitic rocks
11 Granodiorite, monzonite, quartz monzonite, syenite
10 Massive to gneissic quartz diorite, tonalite, trondhjemite
Upper Supergroup
9 9a* Timiskaming Group, 9b** Destor-Porcupine Complex
8 8a, 8n, Blake River Group, 8c*** Blake River (Upper Fm., Tisdale Group)
*a refers to Kirkland Lake Area, south limb of synclinor
**b refers to Kirkland Lake Area, north limb of synclino
***c refers to Timmins Area (Pyke, 1980).
**** (Goodwin, 1965).

Figure 2: Generalized geology of the Timmons-Kirkland Lak







LEGEND

Proterozoic

Keeweenawan diabase (not shown)
12 Cobalt Group

Archean

Matachewan diabase (not shown)

Granitic rocks
11 Granodiorite, monzonite, quartz monzonite, syenite
10 Massive to gneissic quartz diorite, tonalite, trondhjemite

Upper Supergroup
9 9a* Timiskaming Group, 9b** Destor-Porcupine Complex
8 8a, 8n, Blake River Group, 8c*** Blake River (Upper Fm., Tisdale Group)
7 7a, 7b, Kinojevis Group, 7c Kinojevis Group, (Middle Fm., Tisdale Group)
6 6a Larder Lake Group, 6b Stoughton Roquemaure Group, 6c Lower Fm., Tisdale Group
5 5c Porcupine Group

Lower Supergroups
4 4a Skead Group, 4b Hunter Mine Group, 4c Upper Fm., Deloro Group
3 3a Catherine Group, 3c Middle Fm., Deloro Group
2 2a Wabewawa Group, 2c Lower Fm. Deloro Group
1 1a Pacaud tuffs****

*a refers to Kirkland Lake Area, south limb of synclinorium (Jensen 1978c, 1979).
**b refers to Kirkland Lake Area, north limb of synclinorium (Jensen 1976, 1978b).
***c refers to Timmins Area (Pyke, 1980).
**** (Goodwin, 1965).

Figure 2: Generalized geology of the Timmons-Kirkland Lake area.

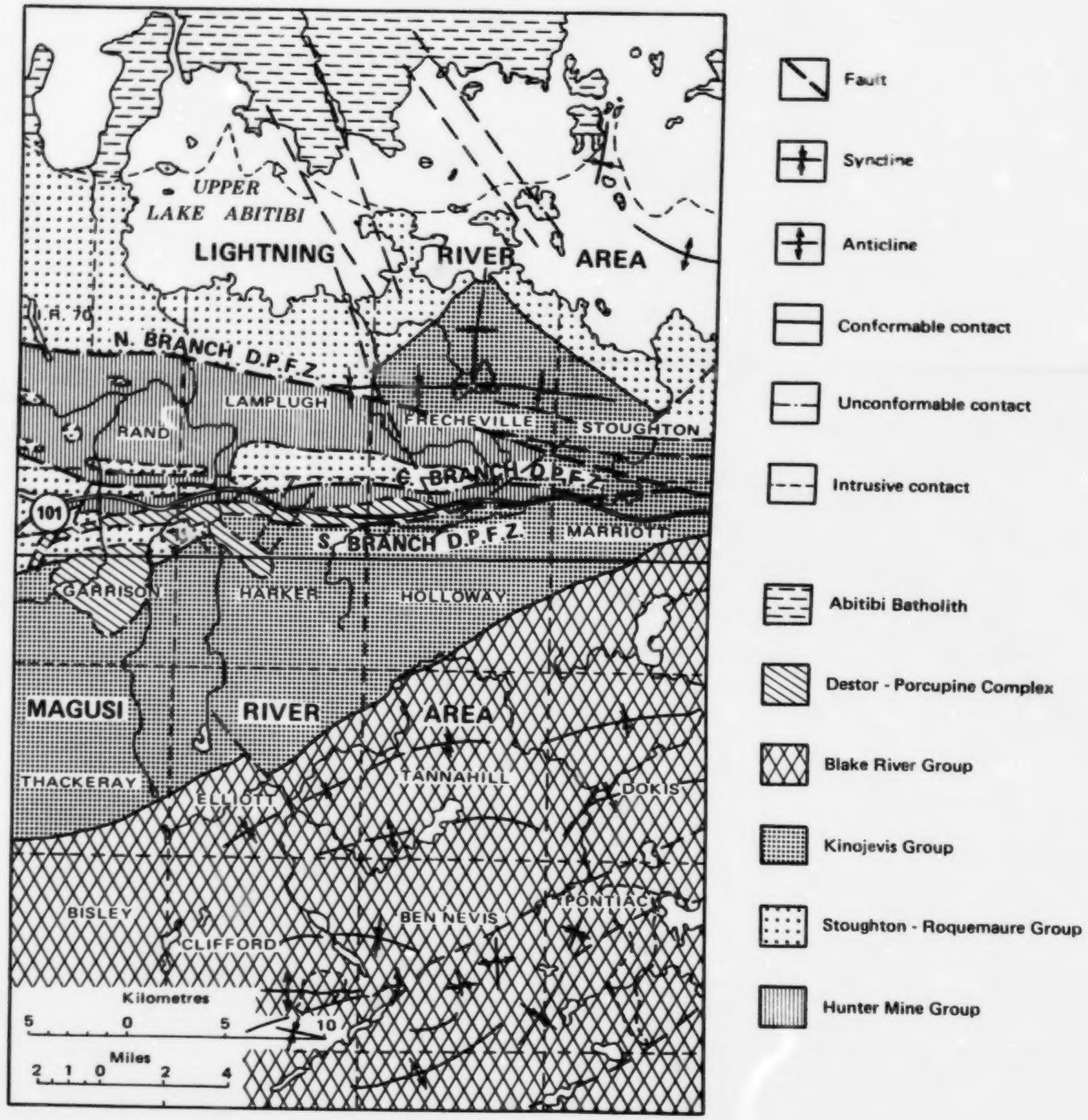
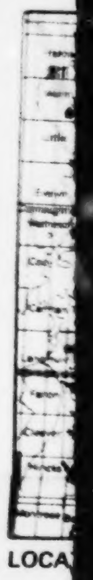
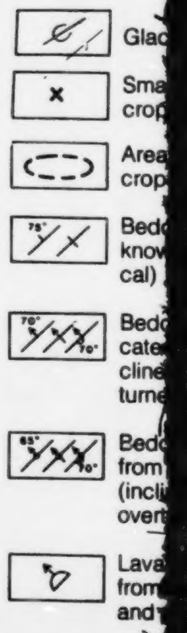


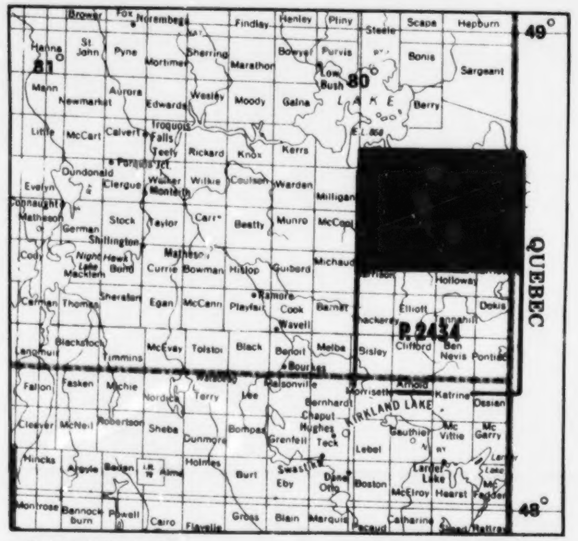
Figure 1: Geological map of the Magusi River and Lightning River areas.





ning River areas.

- ⧅ Fault
- ✛ Syncline
- ✛ Anticline
- ⊟ Conformable contact
- ⊟ Unconformable contact
- ⊟ Intrusive contact
- ▤ Abitibi Batholith
- ▧ Destor - Porcupine Complex
- ▩ Blake River Group
- ▦ Kinojevis Group
- ⁙ Stoughton - Roquemaure Group
- ▥ Hunter Mine Group



LOCATION MAP

Scale 1: 1 584 000
or 1 inch to 25 miles

METAL AND MINERAL ABBREVIATIONS

asb	Asbestos
Au	Gold
cp	Chalcopyrite
IF	Iron Formation
py	Pyrite
qcv	Quartz carbonate vein
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SOURCES OF INFORMATION

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Magnetic declination 10.00°W in 1973.

SYMBOLS

- Glacial striae
- × Small bedrock outcrop
- Area of bedrock outcrop
- Bedding, top unknown; (inclined, vertical)
- Bedding, top indicated by arrow; (inclined, vertical, overturned)
- Bedding, top (arrow) from grain gradation; (inclined, vertical, overturned)
- Lava flow, top (arrow) from pillows shape and packing
- Schistosity; (horizontal, inclined, vertical)
- Gneissosity; (horizontal, inclined, vertical)
- Lineation with plunge
- Geological boundary, observed
- Geological boundary, position interpreted
- Fault; (observed, assumed). Spot indicates downthrow side, arrows indicate horizontal movement.
- Jointing; (horizontal, inclined, vertical)
- Anticline, syncline, with plunge

CREDITS

Geology by L.S. Jensen and assistants, 1973.

Every possible effort has been made to ensure the accuracy of the information presented on this map; however, the Ontario Ministry of Natural Resources does not assume any liability for errors that may occur. Users may wish to verify critical information; sources include both the references listed here, and information on file at the Resident or Regional Geologist's office and the Mining Recorder's office nearest the map-area.

Issued 1982

Information from this publication may be quoted if credit is given. It is recommended that reference be made in the following form:

Jensen, L.S.
1982: Precambrian Geology of the Lightning River Area, Cochrane District; Ontario Geological Survey, Map P.2433, Geological Series - Preliminary Map, Scale 1: 63 360 or 1 inch to 1 mile. Geology 1973.

ULTRAMAFIC INTRUSIVE ROCKS

10 — 10 Peridotite, dunite, pyroxenite, rodingite

INTRUSIVE CONTACT

UPPER CALC-ALKALIC METAVOLCANICS

Felsic Metavolcanics

9 — 9 Massive flow breccia, tuff-breccia, crystal tuff, feldspar, and quartz feldspar porphyritic rhyolitic and dacitic rocks

Mafic and Intermediate Metavolcanics

8 — 8 Massive and pillowed flows, pillow-breccia, pyroclastic breccia, volcanoclastic, tuff, lapilli-tuff, amygdaloidal, porphyritic feldspar, basaltic and andesitic rocks

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Magnesium-Rich Metavolcanics

5 — 5 Grey to green, massive, pillowed, pillow-breccia, hyaloclastic, variolitic, porphyritic feldspar, amygdaloidal basaltic rocks, and interflow sediments

KOMATIITIC METAVOLCANICS

Basaltic Metavolcanics

4 — 4 Massive, pillowed, pillow-breccia, hyaloclastic, variolitic, spinifex-textured basaltic rock

Ultramafic Metavolcanics

3 — 3 Massive, pillowed, polysutured, spinifex-textured ultramafic metavolcanics

UNCONFORMITY

LOWER CALC-ALKALIC METAVOLCANICS

Felsic Metavolcanics

2 — 2 Tuff-breccia, crystal tuff, tuff, quartz and feldspar porphyritic rhyolitic and dacitic rocks, argillite, chert, and ironstone

Mafic and Intermediate Metavolcanics

1 — 1 Massive, pillowed, pillow-breccia, pyroclastic breccia, tuff, amygdaloidal basaltic and andesitic rocks

NOTES

This Legend is used for both Lightning River Area (P.2433) and Magusi Lake Area (P.2434). Not all map-units are present on this map-sheet.

MARGINAL NOTES

LOCATION AND ACCESS

The Magusi River area encompasses the Townships of Bisley, Clifford, Ben Nevis, Pontiac, Thackeray, Elliott, Tannahill, and Dokis, as well as parts of Morrisette, Arnold, Katrine, Ossian, Garrison, Harker, Holloway, and Marriott Townships. The centre of the area is about 25 km northeast of Kirkland Lake.
The southern half of the area is accessible from the south by roads into Bisley, Clifford, Ben Nevis, and southern Pontiac Township from Highway 66 from Kirkland Lake. The northern half is accessible from the north by logging roads from Highway 101 from Matheson. Roads extend south into Thackeray, Garrison, Harker, and northwest Elliott Township. Float equipped aircraft can be used on McDiarmid Lake, Webster Lake, Clarice Lake, Verna Lake, Keith Lake, and Lauramay Lake.

MINERAL EXPLORATION

Mineral exploration in the area has been summarized by Jensen 1972, 1975a, 1975b, 1978. Additional exploration has been carried out on the Canagau Mines Property in Ben Nevis Township and the area southeast of the granitic stock in Clifford Township (Resident Geologist's Assessment files, Ontario Ministry of Natural Resources, Kirkland Lake).

GENERAL GEOLOGY AND STRATIGRAPHY

All the bedrock is of Early Precambrian (Archean) age. The oldest rocks are tholeiitic lavas belonging to the Kinojevis Group. They occur south of the Destor-Porcupine Fault Zone in the Lightning River area and extend south into the area. These rocks form a south-southeast dipping monoclinal succession composed of magnesium-rich and iron-rich tholeiitic basalt alternating as formations within the succession. Tholeiitic dacite and rhyolite form additional units in the succession toward its top. Total thickness of the Kinojevis Group is about 10 km and represents a suite of strongly fractionated mantle-derived magmas showing iron-enrichment.

Overlying the Kinojevis Group is a group of aluminum-enriched lavas referred to as the Blake River Group. This group consists of calc-alkalic basalt, andesite, dacite, and rhyolite as well as some magnesium-rich tholeiitic basalt. The calc-alkalic succession at its base interfingers with the underlying iron-enriched tholeiitic lavas of the Kinojevis Group.

Cutting the calc-alkalic lavas are numerous stocks and sills of gabbro, quartz gabbro, diorite, and quartz-diorite which were likely feeders for the lava flows. Massive, subvolcanic quartz-porphyry domes occur in Clifford and Pontiac Townships, with smaller ones occurring in Ben Nevis and Dokis Townships. Some of these domes were the sources of the nearby rhyolitic felsic tuffs and tuff breccias. The thickness of the Blake River Group in this area is about 10 km; this group was probably generated from the partial melting of deeply buried komatiitic and tholeiitic lavas.

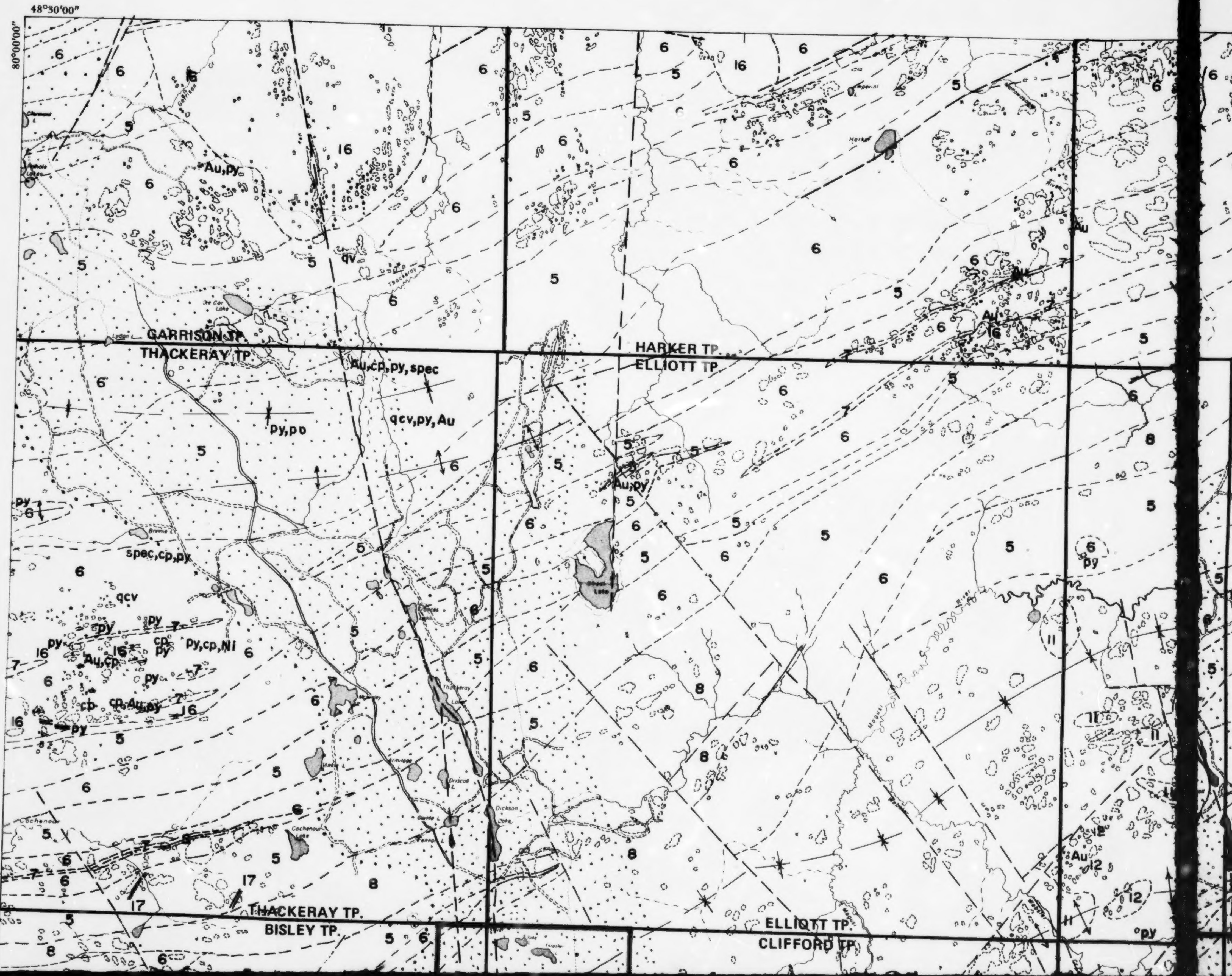
On the north edge of the area, are quartz monzonite, syenite, and syenodiorite stocks and dikes. These intrusions become progressively potash-enriched, smaller, and less numerous southward.

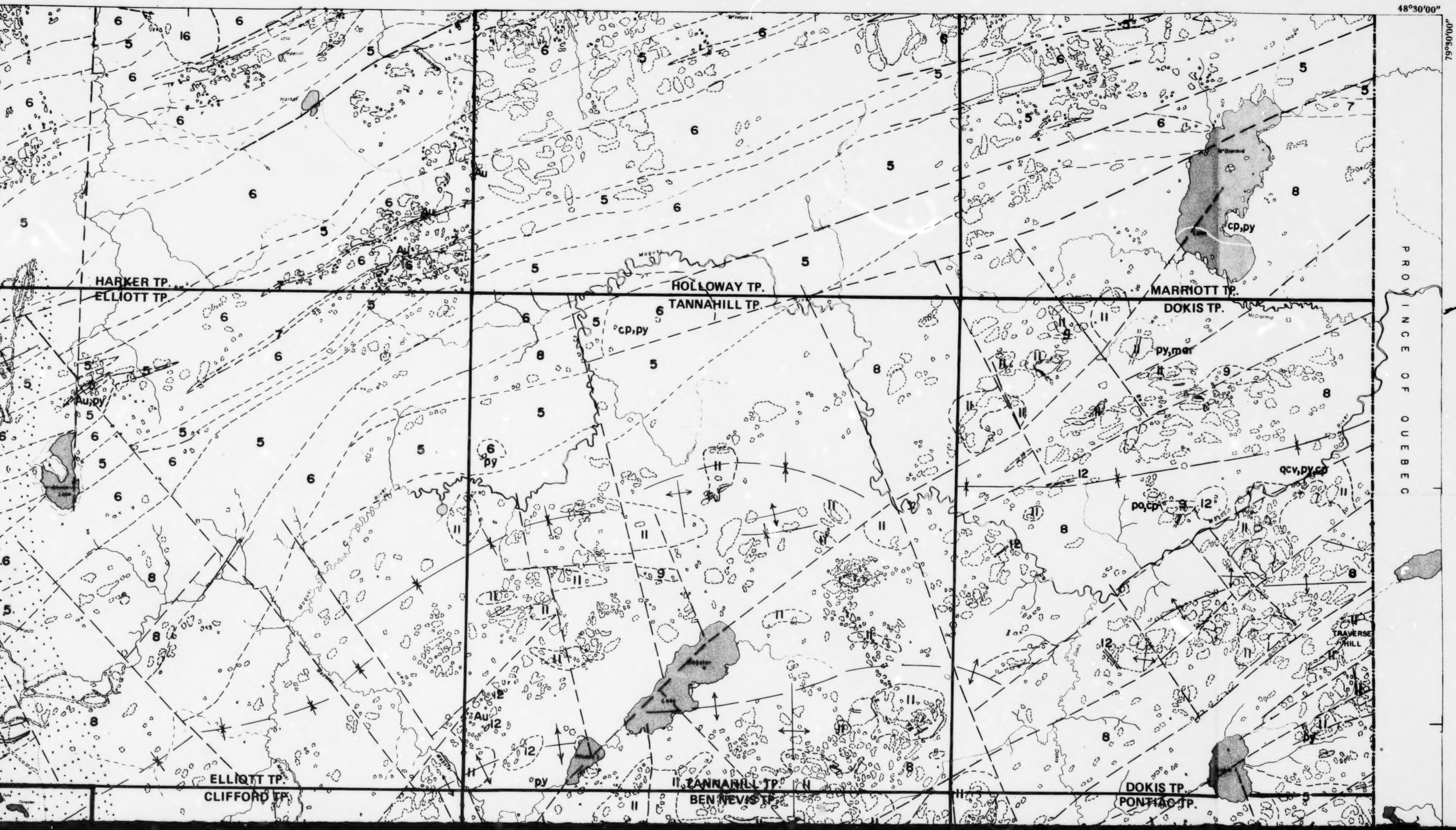
The generalized geology of this map-area and the adjacent Lightning River map-area is shown in Figure 1. Sample locations of metavolcanics for which whole rock analyses are available are shown in Figure 2.

STRUCTURAL GEOLOGY

The volcanic rocks are contained in an east-plunging synclinorium. The lower Kinojevis Group is a steeply dipping south-southeast facing monoclinal succession, the base of which is cut off by the Destor-Porcupine Fault Zone farther north. Considerable subsidence of these lavas has occurred along the south margin of the fault zone as well as by further subsidence by tilting south. Present subgreenschist metamorphic grades in the lower rocks suggest the lavas were not deeply buried by succeeding lavas but instead were tilted southward as succeeding deposited lavas were being progressively confined southward. This would also explain why subvolcanic feeders for the tholeiitic lavas are not detected.

The Blake River Group is folded into a series of concentric synclines and anticlines curvilinear around the




48°30'00"
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HARKER TP.
ELLIOTT TP.
HOLLOWAY TP.
TANNAHILL TP.
MARRIOTT TP.
DOKIS TP.
ELLIOTT TP.
CLIFFORD TP.
TANNAHILL TP.
BEN NEVIS TP.
DOKIS TP.
PONTIAC TP.
PROVINCE OF QUEBEC
TRAVERSE HILL
Au
Au
Au
Au,py
Au
cp,py
cp,py
py
py,mar
qcv,py,cp
po,cp
py
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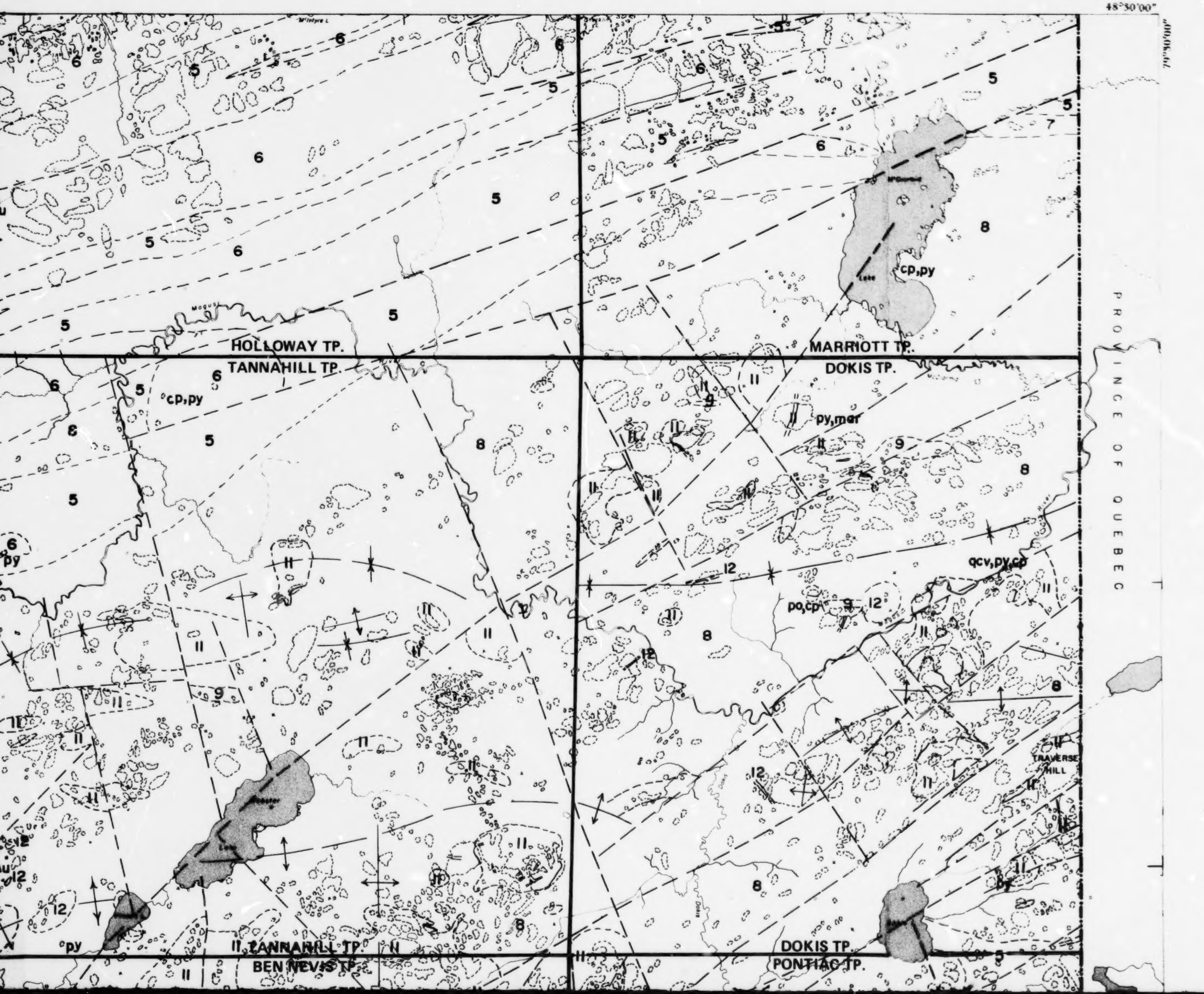

Ministry of Natural Resources
Ontario

Hon. Alan W. Pope
Minister

W. T. Foster
Deputy Minister

ONTARIO GEOLOGICAL SURVEY

MAP P.2434

GEOLOGICAL SERIES - PRELIMINARY MAP

PRECAMBRIAN GEOLOGY OF THE

MAGUSI RIVER AREA

COCHRANE AND TIMISKAMING DISTRICTS

Scale 1:63 360 or 1 inch to 1 mile

Miles 1 0 1 2 Miles

Kilometres 1 0 1 2 3 Kilometres

NTS Reference 32 D/5
ODM-GSC Aeromagnetic Map 46G
ODM Geological Compilation Map 2205

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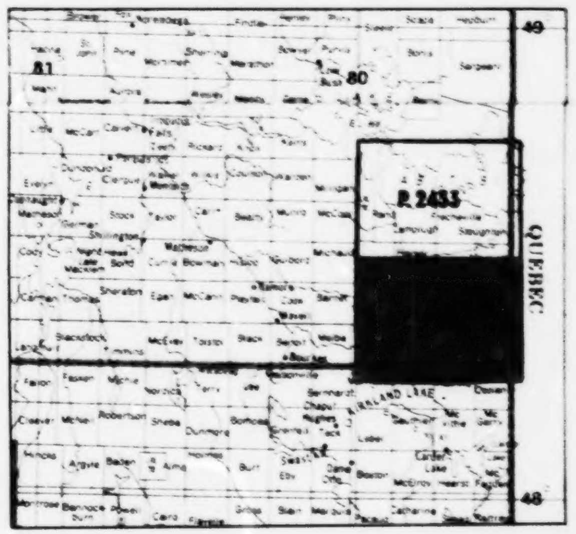


LOCATION MAP

Scale: 1:1 584 000
or 1 inch to 25 miles

LEGEND

PHANEROZOIC
CENOZOIC
QUATERNARY
PLEISTOCENE AND RECENT

rhyolitic felsic tuffs and tuff breccias. The thickness of the Blake River Group in this area is about 10 km; this group was probably generated from the partial melting of deeply buried komatiitic and tholeiitic lavas.

On the north edge of the area, are quartz monzonite, syenite, and syenodiorite stocks and dikes. These intrusions become progressively potash-enriched, smaller, and less numerous southward.

The generalized geology of this map-area and the adjacent Lightning River map-area is shown in Figure 1. Sample locations of metavolcanics for which whole rock analyses are available are shown in Figure 2.

STRUCTURAL GEOLOGY

The volcanic rocks are contained in an east-plunging synclinorium. The lower Kinojevis Group is a steeply dipping south-southeast facing monoclinal succession, the base of which is cut off by the Destor-Porcupine Fault Zone farther north. Considerable subsidence of these lavas has occurred along the south margin of the fault zone as well as by further subsidence by tilting south. Present subgreenschist metamorphic grades in the lower rocks suggest the lavas were not deeply buried by succeeding lavas but instead were tilted southward as succeeding deposited lavas were being progressively confined southward. This would also explain why subvolcanic feeders for the tholeiitic lavas are not detected.

The Blake River Group is folded into a series of concentric synclines and anticlines curvilinear around the rhyolitic domes in Clifford and Pontiac Townships. Folding was caused by the subsidence of the underlying komatiitic and tholeiitic lavas. As the buried rocks sank, partial melting occurred, with the calc-alkalic magmas forming and coming to surface and the remaining garnet-rich residues sinking into the mantle. As the calc-alkalic lavas were deposited, they would begin to collapse into folds around the more bouyant rhyolite domes.

Some of the faults may be related to subsidence but in general the northeast and the northwest striking faults are later isostatic adjustments in the earth's crust.

ECONOMIC GEOLOGY

Gold, copper, zinc, silver, lead, and molybdenum mineralization occurs in the calc-alkalic volcanic rocks, particularly toward the core of the synclinorium. This mineralization occurs mainly in rusty pyrite-rich gossans.

If the calc-alkalic rocks were formed by about 10 percent partial melting of ultramafic and mafic rocks, 10 km^3 of mafic to ultramafic rock would have to be partially melted to form 1 km^3 of calc-alkalic magma. Distilled sulphides would be expected to be channelled to surface by hydrothermal fluids and precipitated as sulphide deposits; thus the greater the amount of calc-alkalic rocks produced, the more chance of stacked *massive sulphide deposits similar to those in the Noranda area, Quebec.*

Gold and copper mineralization is associated with the syenitic intrusions in the north part of the area. The syenitic rocks were probably the result of the partial melting of deeply buried clay-rich carbonaceous sedimentary rocks which were possibly gold-bearing. These sediments were eroded off an earlier formed calc-alkalic volcanic pile to the north in the Lightning River area and were subsequently deeply buried by komatiitic and tholeiitic lavas.

REFERENCES

Gledhill, T.L.
1929:Ben Nevis, Munro, Kamiskotia, and Other Base-Metal Areas, District of Cochrane and Timiskaming; Ontario Department of Mines, 1928, Volume 37, pt.3, p.1-52. Accompanied by Map No.37g, Scale 1 inch to 1 mile.

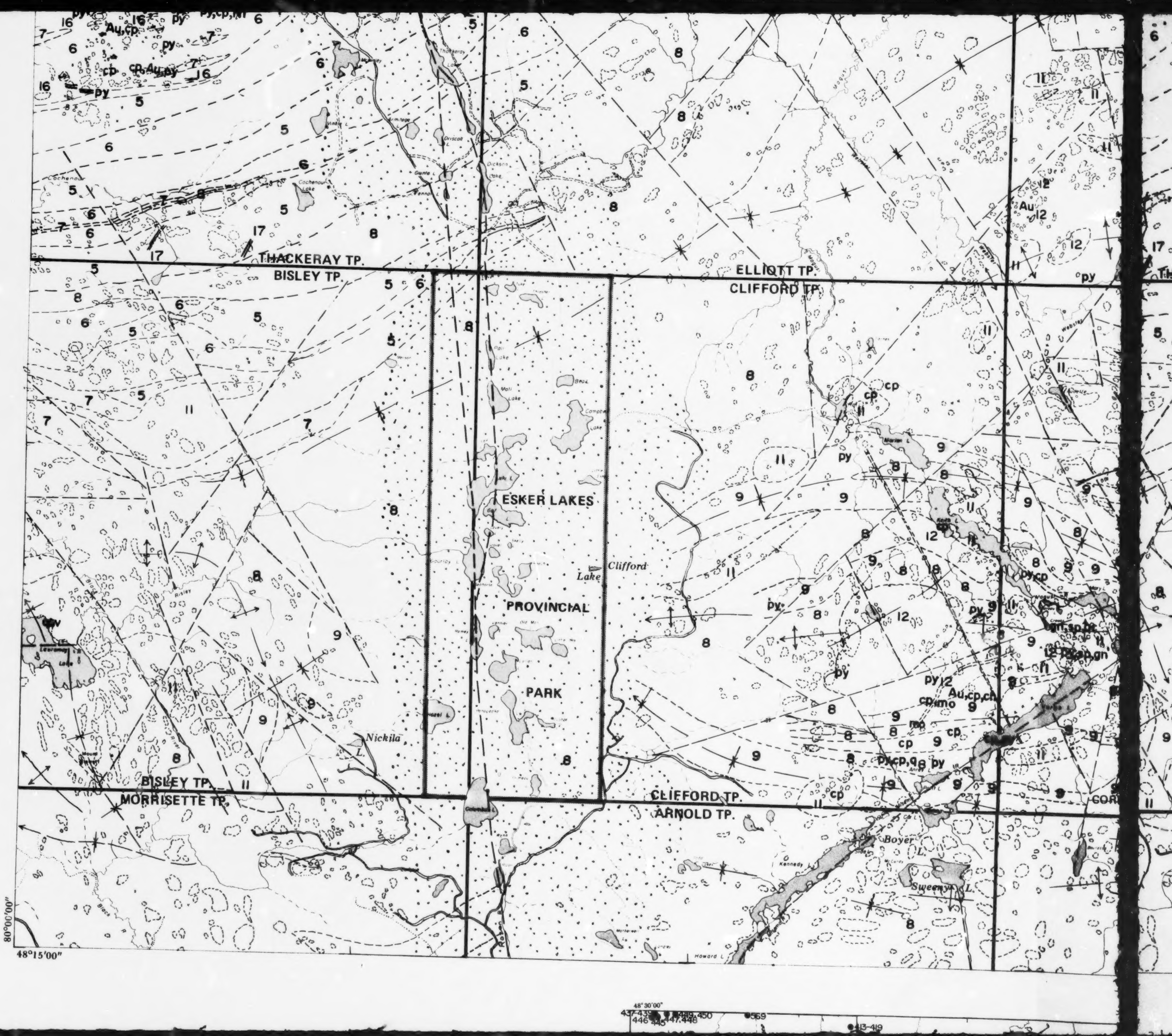
Hogg, W.A.
1964:Arnold and Katrine Townships; Ontario Department of Mines, Geological Report 29, 15p. Accompanied by Map 2061, Scale 1 inch to ½ mile.

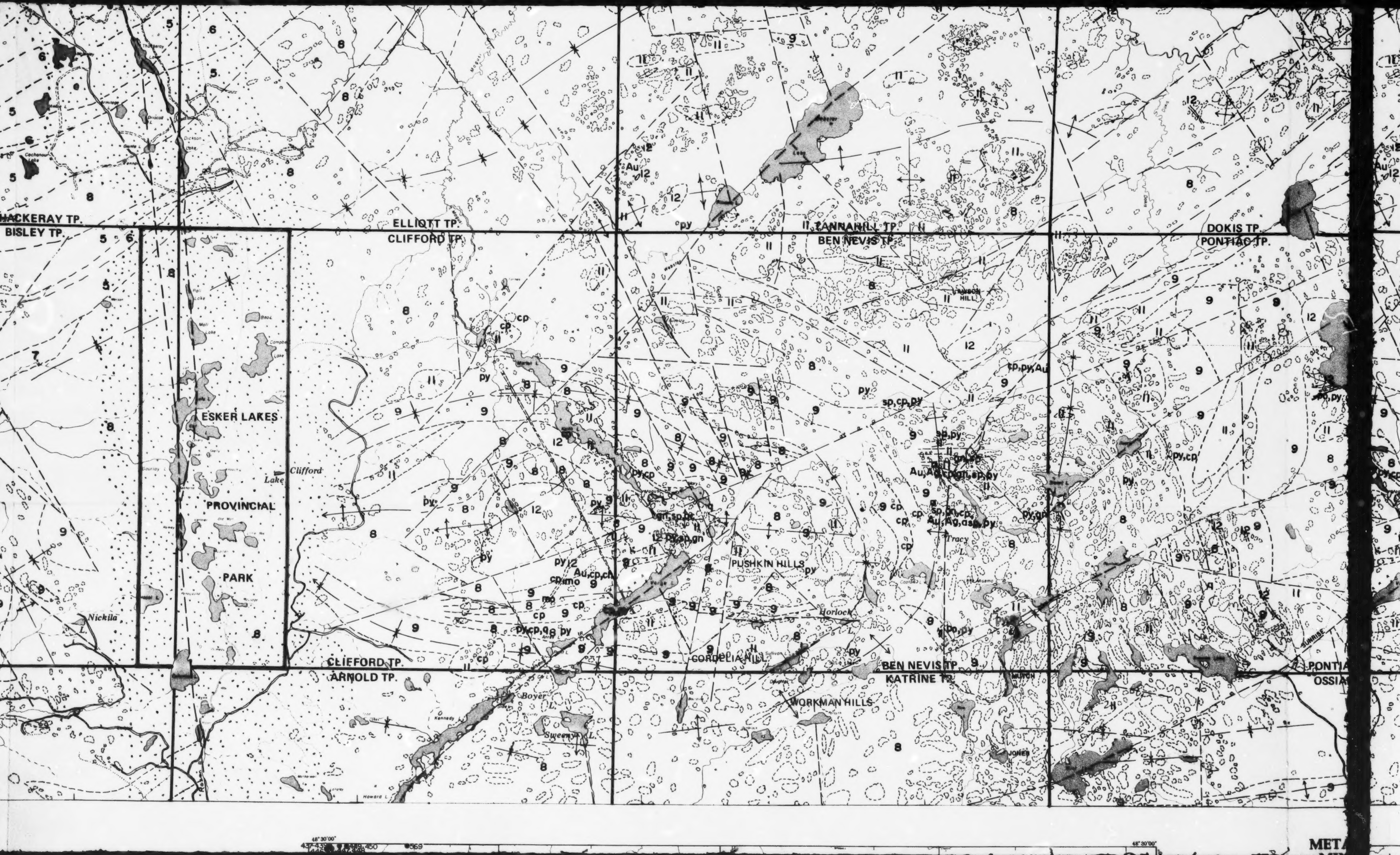
Jensen, L.S.
1972: Geology of Melba and Bisley Townships, District of Timiskaming; Ontario Department of Mines and Northern Affairs, Geological Report 103, 25p. Accompanied by Map 2252, Scale 1 inch to ½ mile.

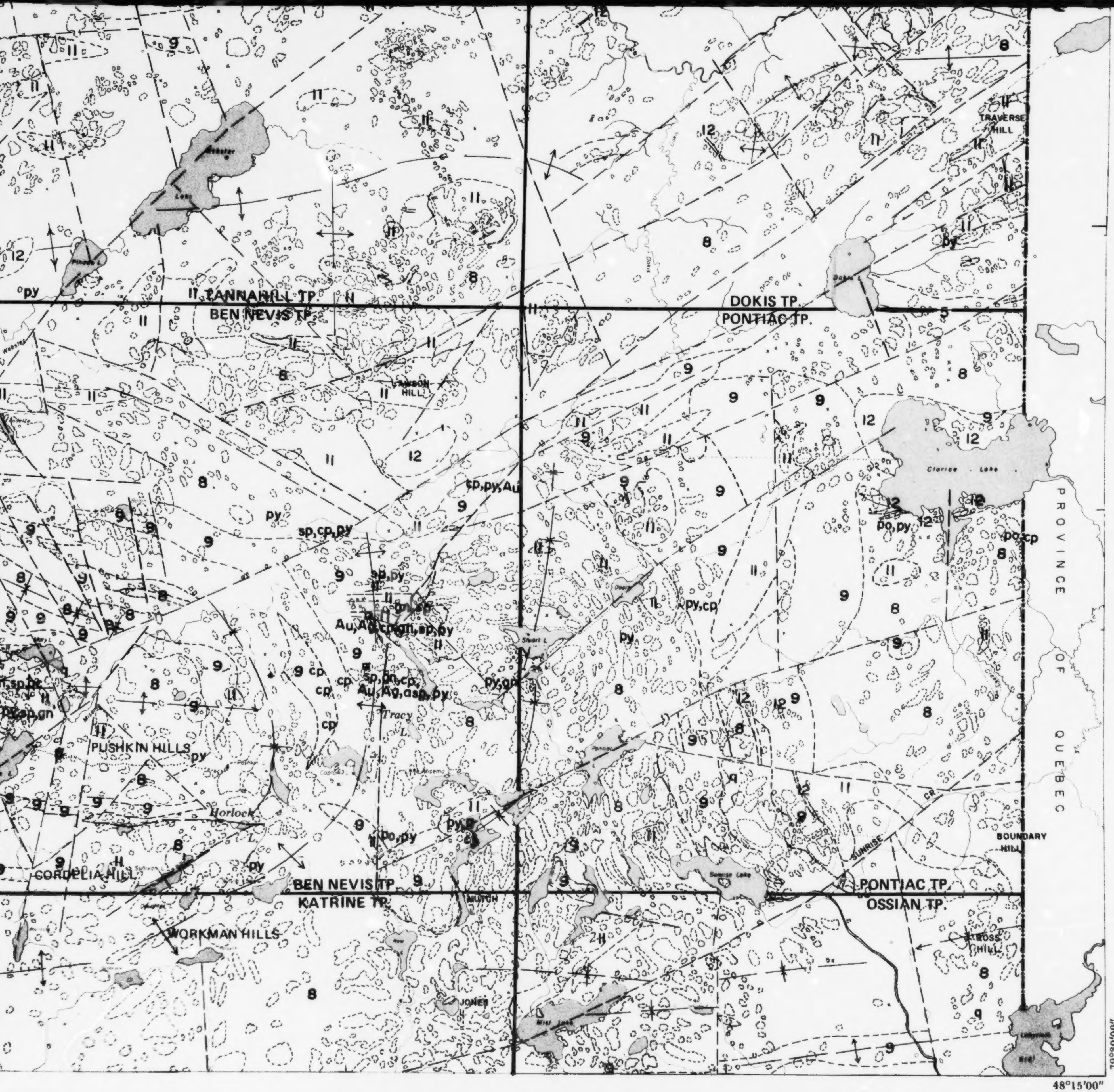
1975a:Pontiac and Ossian Townships, District of Timiskaming and Cochrane; Ontario Division of Mines, Geological Report 125, 41p. Accompanied by Map 2206, Scale 1 inch to ½ mile.

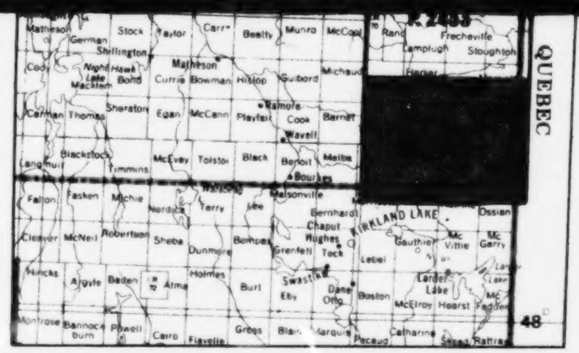
1975b: Clifford and Ben Nevis Townships, District of Cochrane; Ontario Division of Mines, Geological Report 132, 55p. Accompanied by Map 2283, Scale 1 inch to ½ mile.

1978a:Geology of Thackeray, Elliott, Tannahill, and Dokis Townships, District of Cochrane; Ontario Geological Survey, Report 165, 71p. Accompanied by




THACKERAY TP.
BISLEY TP.
ELLIOTT TP.
CLIFFORD TP.
TANNAHILL TP.
BEN NEVIS TP.
DOKIS TP.
PONTIAC TP.
ESKER LAKES
PROVINCIAL
PARK
Clifford
Lake
Nickila
L.
PUSHKIN HILLS
CORDELIA HILL
Horlock
Tracy
L.
CLIFFORD TP.
ARNOLD TP.
BEN NEVIS TP.
KATRINE TP.
WORKMAN HILLS
Boyer
L.
Sweeney
L.
Howard L.
PONTIA
OSSIA
SUNRISE
Au
cp
py
cp,py,Au
sp,cp,py
sp,py
Au,Ag,cp,gn,sp,py
Au,Ag,asp,py
py,cp
py,gn
Au,cp,ch
cp,mo
py,cp,q
gn,sp,py
48°30'00"
META





LOCATION MAP

Scale: 1:1 584 000
or 1 inch to 25 miles

LEGEND

PHANEROZOIC
CENOZOIC
QUATERNARY
PLEISTOCENE AND RECENT
Till, reworked till, esker sand and gravel, varved clay, dune sand, alluvium, and peat

UNCONFORMITY

PRECAMBRIAN
MIDDLE TO LATE PRECAMBRIAN (PROTEROZOIC)
MAFIC INTRUSIVE ROCKS

18 — 18 Diabase, quartz diabase (Keweenawan)

INTRUSIVE CONTACT

EARLY PRECAMBRIAN (ARCHEAN)
MAFIC INTRUSIVE ROCKS

17 — 17 Diabase (Matachewan)

INTRUSIVE CONTACT

FELSIC INTRUSIVE ROCKS

16 — 16 Equigranular and porphyritic syenodiorite, monzonite, feldspar porphyry, pegmatite, lamprophyre

INTRUSIVE CONTACT

ALKALIC METAVOLCANICS
Felsic Metavolcanics

15 — 15 Sodic trachyte, benmorite, alkalic dacite, rhyolite

Mafic Metavolcanics

14 — 14 Hawaiite, alkali basalt, mugearite, nephelinite

METASEDIMENTS

13 — 13 Conglomerate, wacke, arkose, argillite, sandstone, ironstone

UNCONFORMITY

FELSIC INTRUSIVE ROCKS

12 — 12 Trondhjemite, quartz diorite, diorite

INTRUSIVE CONTACT

MAFIC INTRUSIVE ROCKS

11 — 11 Gabbro, quartz gabbro, hornblende gabbro, granophyre

INTRUSIVE CONTACT

ULTRAMAFIC INTRUSIVE ROCKS

10 — 10 Peridotite, dunite,

METAL AND

Report 29, 15p. Accompanied by Map 2061, Scale 1 inch to ½ mile.

Jensen, L.S.

1972: Geology of Melba and Bisley Townships, District of Timiskaming; Ontario Department of Mines and Northern Affairs, Geological Report 103, 25p. Accompanied by Map 2252, Scale 1 inch to ½ mile.

1975a:Pontiac and Ossian Townships, District of Timiskaming and Cochrane; Ontario Division of Mines, Geological Report 125, 41p. Accompanied by Map 2206, Scale 1 inch to ½ mile.

1975b: Clifford and Ben Nevis Townships, District of Cochrane; Ontario Division of Mines, Geological Report 132, 55p. Accompanied by Map 2283, Scale 1 inch to ½ mile.

1978a:Geology of Thackeray, Elliott, Tannahill, and Dokis Townships, District of Cochrane; Ontario Geological Survey, Report 165, 71p. Accompanied by Maps 2367 and 2368, Scale 1 inch to ½ mile.

1978b:Geology of Marriott and Stoughton Townships, District of Cochrane; Ontario Geological Survey, Report 173, 73p. Accompanied by Map 2390, Scale 1 inch to ½ mile.

1981:A Petrogenic Model for the Archean Abitibi Belt in the Kirkland Lake Area, Ontario; Unpublished Ph.D. Thesis, University of Saskatchewan, Canada, 519p.

1982:Lightning River Area, Cochrane District; Ontario Geological Survey, Map P.2433, Geological Series-Preliminary Map, Scale 1:63 300 or 1 inch to 1 mile.

Rupert, R.J., and Lovell, H.H.

.1970: Geology of Bernhardt and Morrisette Townships, District of Timiskaming; Ontario Department of Mines, Geological Report 84, 27p. Accompanied by Map 2193, Scale 1 inch to ½ mile.

Satterly, J.

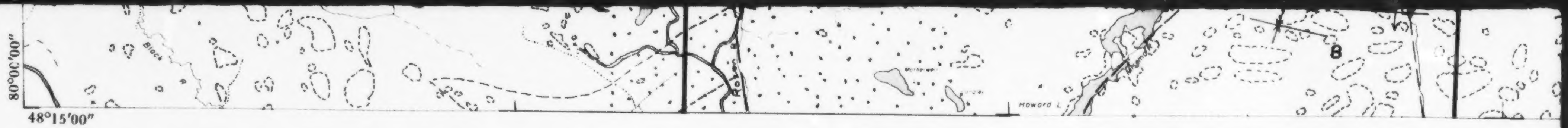
1949: Geology of Garrison Township; Ontario Department of Mines, Vol. 58, pt.4, p.1-33. Accompanied by Map No. 1949-1, Scale 1 inch to 1000 feet.

1952:Geology of Harker Township, Ontario Department of Mines, 1951, Vol. 60, pt.7, p.1-47. Accompanied by Map 1951-4, Scale 1 inch to 1000 feet.

1954:Geology of the North Half of Holloway Township; Ontario Department of Mines, 1953, Volume 62, pt.7, p.1-38. Accompanied by Map No. 1953-4, Scale 1 inch to 1000 feet.

Wolfe, W.J.

1977:Early Precambrian Volcanogenic Sulphide Mineralization in Ben Nevis Township, District of Cochrane; Ontario Geological Survey, Study 19, 39p.



SYMBOLS

×	Small bedrock outcrop
( )	Area of bedrock outcrop
↗ 30°	Lineation with plunge
⁀	Geological boundary, observed
‐ ‐	Geological boundary, position interpreted
	Fault; (observed, assumed). Spot indicates down throw side, arrows indicate horizontal movement.
70° 70°	Anticline, syncline with plunge
◪ 150'	Shaft

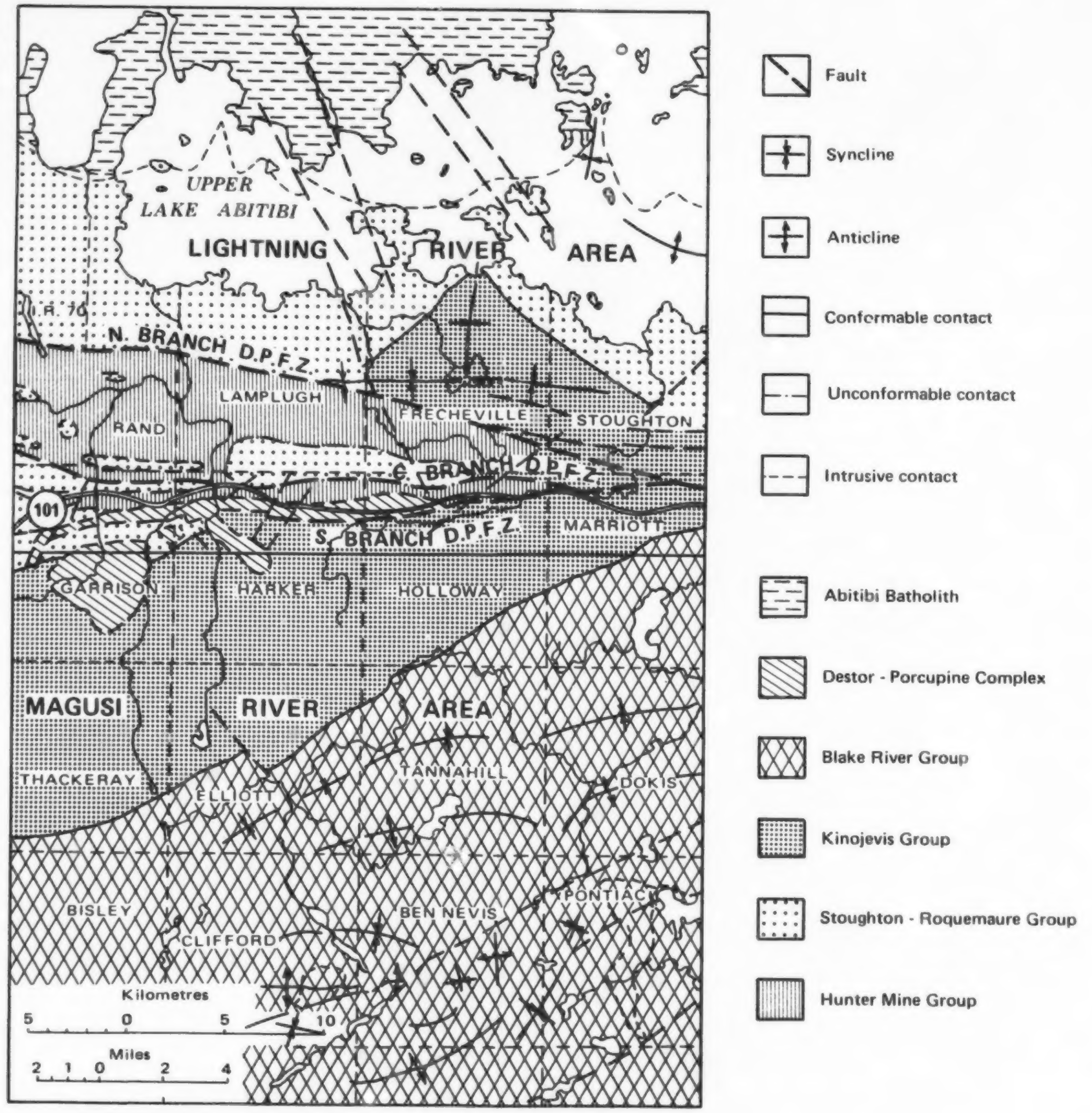


Figure 1: Geological map of the Magusi River and Lightning River areas.

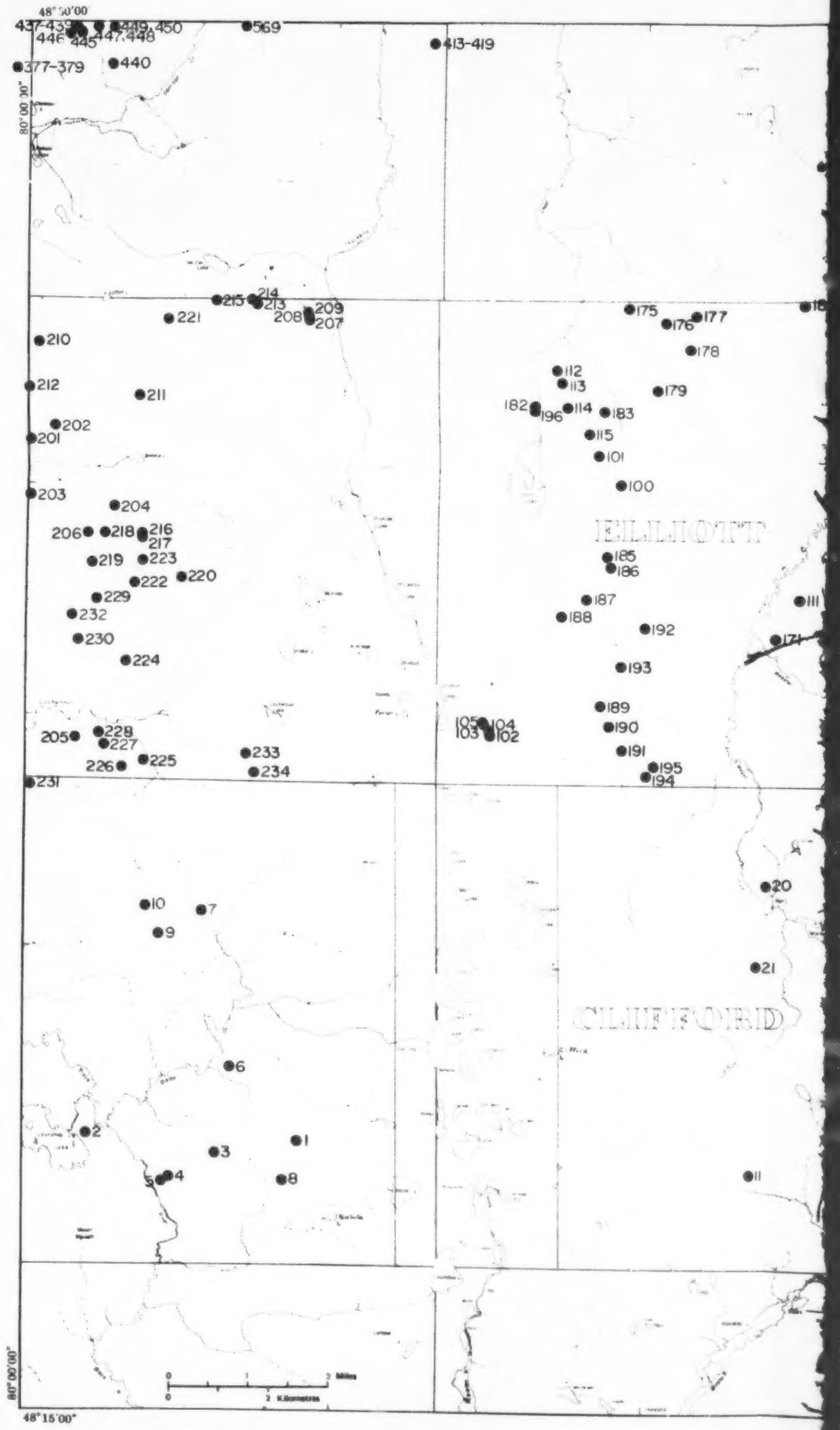


Figure 2: Sample locations for chemical analyses.

- Fault
- Syncline
- Anticline
- Conformable contact
- Unconformable contact
- Intrusive contact
- Abitibi Batholith
- Destor - Porcupine Complex
- Blake River Group
- Kinojevis Group
- Stoughton - Roquemaure Group
- Hunter Mine Group

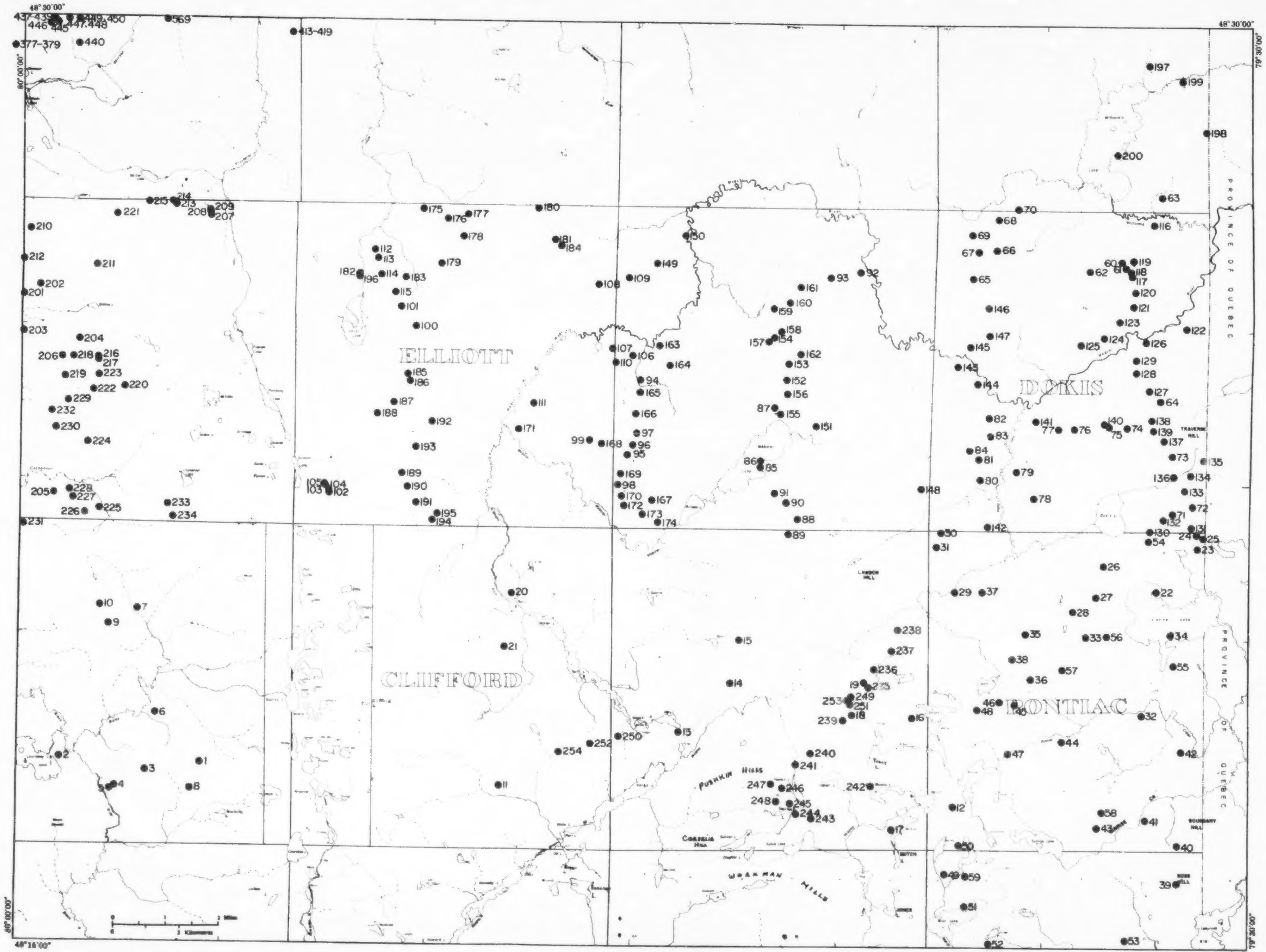


Figure 2: Sample locations for chemical analyses.

METAL AND MINERAL ABBREVIATIONS

Ag	Silver	**po**	Pyrr
asp	Arsenopyrite	**py**	P
Au	Gold	**q**	Q
ch	Chert	**qc**	Quartz carbo
cp	Chalcopyrite	**qcv**	Quartz carbonate
gn	Galena	**qv**	Quartz
mar	Marcasite	**sp**	Spha
mo	Molybdenite	**spec**	Specu
Ni	Nickel		

SOURCES OF INFORMATION

Base map derived from Forest Resources Inventory maps, Land Waters Group, Ontario Ministry of Natural Resources.

Geology not tied to survey lines.

Magnetic declination 10.00°W, 1968-1972.

CREDITS

Geology of Bisley, Clifford, Ben Nevis, Pontiac, Ossian, Thackeray, Tannahill, Dokis, and Mariott Townships by L.S. Jensen and assist 1968, 1969, 1970, 1971, 1972.

Geology of Garrison and Harker Townships by J. Satterly (1949, 195

Geology of Morisette Township by R.J. Rupert and H.L. Lovell, 1967

Geology of Arnold and Katrine Townships by W.A. Hogg, 1963.

Every possible effort has been made to ensure the accuracy of the mation presented on this map; however, the Ontario Ministry of N Resources does not assume any liablility for errors that may occur. may wish to verifty critical information; sources include both the ence listed here, and information on file at the Resident or Regiona ologist's office and the Mining Recorder's office nearest the map-ar

Issued 1982

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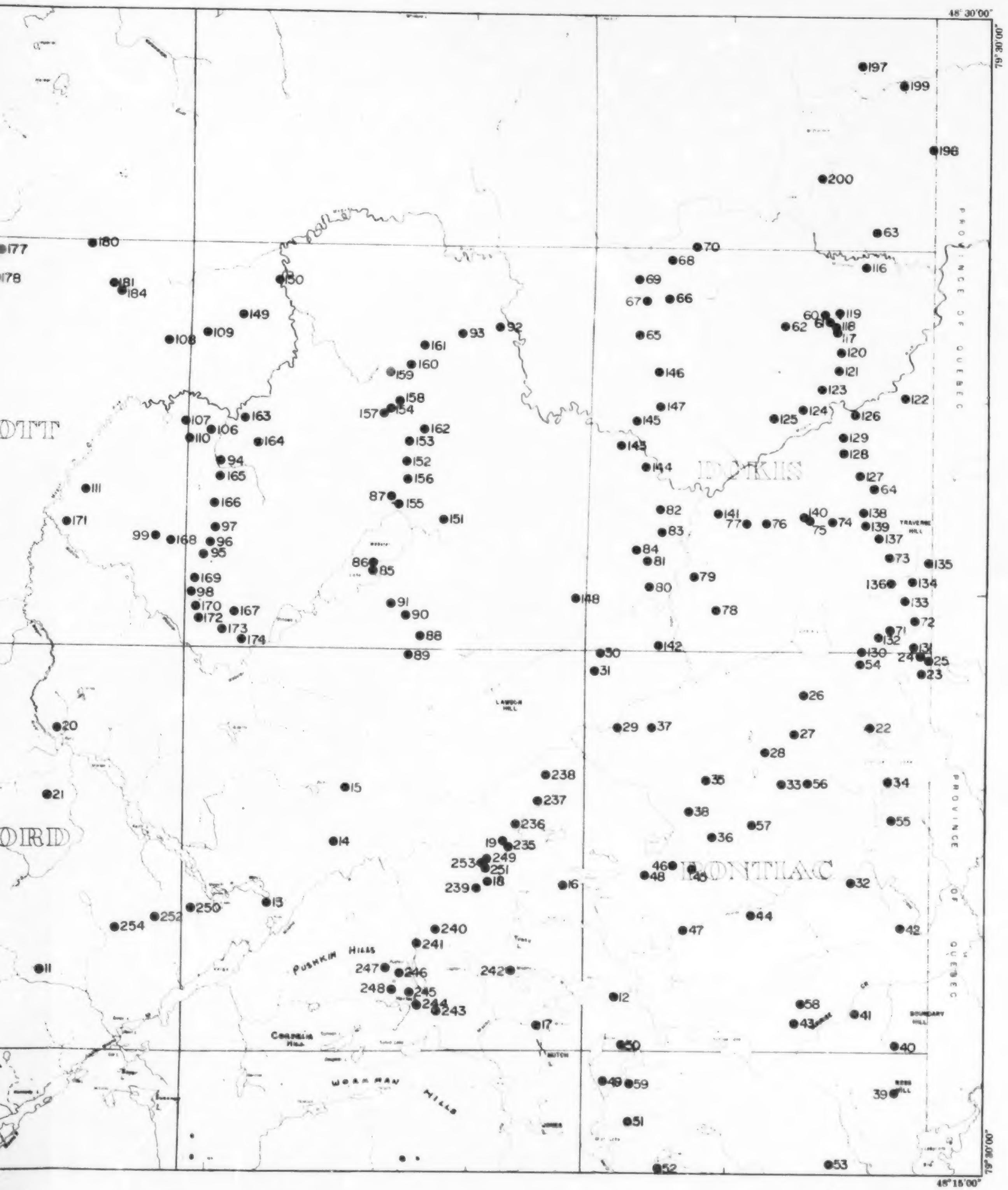
1982: Precambrian Geology of the Magusi River Area, Cochran Timiskaming Districts; Ontario Geological Survey, Map P. Geological Series-Preliminary Map, Scale 1:63 360 or 1 inc mile. Geology 1968-1972.



METAL AND MINERAL ABBREVIATIONS

Ag	Silver	**po**	Pyrrhotite
asp	Arsenopyrite	**py**	Pyrite
Au	Gold	**q**	Quartz
ch	Chert	**qc**	Quartz carbonate
cp	Chalcopyrite	**qcv**	Quartz carbonate vein
gn	Galena	**qv**	Quartz vein
mar	Marcasite	**sp**	Sphalerite
mo	Molybdenite	**spec**	Specularite
Ni	Nickel		

SOURCES OF INFORMATION

Base map derived from Forest Resources Inventory maps, Lands and Waters Group, Ontario Ministry of Natural Resources.

Geology not tied to survey lines.

Magnetic declination 10.00°W, 1968-1972.

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Geology of Garrison and Harker Townships by J. Satterly (1949, 1951).

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1982: Precambrian Geology of the Magusi River Area, Cochrane and Timiskaming Districts; Ontario Geological Survey, Map P.2434, Geological Series-Preliminary Map, Scale 1:63 360 or 1 inch to 1 mile. Geology 1968-1972.

INTRUSIVE CONTACT

MAFIC INTRUSIVE ROCKS

11 | 11 Gabbro, quartz gabbro, hornblende gabbro, granophyre

INTRUSIVE CONTACT

ULTRAMAFIC INTRUSIVE ROCKS

10 | 10 Peridotite, dunite, pyroxenite, rodingite

INTRUSIVE CONTACT

UPPER CALC-ALKALIC METAVOLCANICS

Felsic Metavolcanics

9 | 9 Massive flow breccia, tuff-breccia, crystal tuff, feldspar, and quartz feldspar porphyritic rhyolitic and dacitic rocks

Mafic and Intermediate Metavolcanics

8 | 8 Massive and pillowed flows, pillow-breccia, pyroclastic breccia, volcanoclastic, tuff, lapilli-tuff, amygdaloidal, porphyritic feldspar, basaltic and andesitic rocks

THOLEIITIC METAVOLCANICS

Felsic Metavolcanics

7 | 7 Spherulitic and granular tuff, tuff-breccia, cherty tuff, dacite, and rhyolite

Iron-Rich Mafic Metavolcanics

6 | 6 Black to dark green, massive, pillowed, pillow-breccia, hyaloclastic, variolitic, amygdaloidal basalt, andesite rocks, and interflow sediments

Magnesium-Rich Metavolcanics

5 | 5 Grey to green, massive, pillowed, pillow-breccia, hyaloclastic, variolitic, porphyritic feldspar, amygdaloidal basaltic rocks, and interflow sediments

KOMATIITIC METAVOLCANICS

Basaltic Metavolcanics

4 | 4 Massive, pillowed, pillow-breccia, hyaloclastic, variolitic, spinifex-textured basaltic rock

Ultramafic Metavolcanics

3 | 3 Massive, pillowed, polysutured, spinifex-textured ultramafic metavolcanics

UNCONFORMITY

LOWER CALC-ALKALIC METAVOLCANICS

Felsic Metavolcanics

2 | 2 Tuff-breccia, crystal tuff, tuff, quartz and feldspar porphyritic rhyolitic and dacitic rocks, argillite, chert, and ironstone

Mafic and Intermediate Metavolcanics

1 | 1 Massive, pillowed, pillow-breccia, pyroclastic breccia, tuff, amygdaloidal basaltic and andesitic rocks

NOTES

This Legend is used for both Lightning River Area (P.2433) and Magusi Lake Area (P.2434). Not all map-units are present on this map-sheet.